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Empirical modeling of late quaternary sediment supply to the Gulf of Mexico shelf margin in a cool lowstand world

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EMPIRICAL MODELING OF LATE QUATERNARY SEDIMENT SUPPLY TO THE GULF OF MEXICO SHELF MARGIN IN A COOL LOWSTAND WORLD

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ABSTRACT

During the last glacial period, river systems responded to the lowering of sea level by incising valleys, extending across the continental shelf, and depositing deltas on the shelf margin and upper slope of surrounding basins. These deltas represent important repositories for sediment delivered to the coastal ocean and provide partial records of sediment discharged from continental landmasses. Over relatively short time scales, like that of the last glacial-interglacial cycle, hinterland drainage areas, relief and lithology can be considered constants while the primary variable of significance is climate change (Syvitski and Milliman, 2007). It is reasonable to assume, therefore, that supply rates would have been different during the last glacial period, relative to the modern interglacial.

For the northern Gulf of Mexico (GOM), a recently summarized body of data documents volumes of sediment sequestered in a variety of shelf and shelf-margin deltas (Anderson and Fillon, 2004): from this database, individual workers have suggested the sediment discharge of large rivers was significantly greater during the last glacial period than the present. This thesis tests the plausibility and necessity of these previous interpretations and attempts to move towards a first-generation sediment budget that contrasts interglacial versus glacial periods for river systems that discharge to the northern GOM. Recently developed empirical models of sediment supply (Syvitski et al., 2003; Syvitski and Milliman, 2007) are used to estimate supply rates for modern river systems and the role of glacial-period boundary conditions on supply rates. The spectrum of possible sediment fluxes derived is then used to ascertain changes in sediment supply necessary to account for the observed sediment mass in shelf-margin deltas.

This thesis has found that for the last glacial period: (a) model estimates for sediment yield are lower than present values due to temperature depressions; (b) model estimates for

sediment discharge are lower, except for the Mobile and Brazos Rivers due to their increase in drainage area; and (c) in all but one case, observed sediment volumes are significantly less than model estimates of sediment discharge over the inferred time periods of deposition.

INTRODUCTION

During the last glacial period, river systems around the world responded to the lowering of sea level by incising valleys, extending across the subaerial continental shelf, and depositing deltas on the shelf margin and upper slope of surrounding basins. Deltas located at shelf margins represent an important repository for sediment delivered to the coastal ocean during glacial periods and provide a partial record of sediment discharged from continental landmasses. In contrast to sediment stored elsewhere in a dispersal system, sediment sequestered in shelf-margin deltas can be measured using high-resolution seismic data, which has become widely available over the last 2-3 decades, and provides a basis for initial development of sediment budgets, and how they change over time in response to various forcing mechanisms.

Controls on rates of fluvial-sediment supply to a basin margin are reasonably well known and include drainage area, relief, lithology, and climate (Syvitski and Milliman, 2007). Changes in controls on sediment supply should have a noticeable impact on changes in rates of sediment delivery to depositional basins over time. Over relatively short time scales, like that of the last 100-kyr glacial-interglacial cycle, hinterland drainage areas, relief and lithology are constants, with the primary variable of significance being climatic change. It is reasonable to assume, therefore, that supply rates would have been different during the last glacial period, relative to the modern interglacial. However, the ability to unravel relationships between climatic controls and sediment supply is commonly limited by observations and geochronologic data.

For the northern Gulf of Mexico (GOM), a recently summarized body of data documents volumes of sediment sequestered in a variety of shelf and shelf-margin deltas (Anderson and Fillon, 2004): from this database, individual workers have suggested the sediment discharge of large rivers was significantly greater during the last glacial period than today, but the impacts of climate changes on sediment supply during the last glacial to interglacial cycle were overshadowed by the effect of sea-level fall (Abdulah et al., 2004; Anderson et al., 2004; Banfield and Anderson, 2004). If correct, such interpretations of higher sediment supply during a cool glacial period would contradict general relationships that have been established in the literature between climate variables and sediment supply (e.g. Summerfield and Hulton, 1994; Milliman and Syvitski, 1992; Mulder and Syvitski, 1996; Hovius, 1998; Leeder et al., 1998; Harrison, 2000; Syvitski et al., 2003).

Published interpretations of sediment volumes in shelf-margin deltas from the last glacial period in the northern GOM provide a starting point to begin to address the complex relationships between sediment flux and climate change over the last glacial to interglacial cycle. This thesis tests the plausibility and necessity of previous interpretations of greater sediment supply during the last glacial period, as compared with today through applications of models. In other words, this thesis asks the following two questions. First, are higher sediment supply rates necessary to explain observed sediment volumes? Second, are higher sediment supply rates plausible with glacial boundary conditions? This approach represents a move towards a first-generation sediment budget that contrasts interglacial vs. glacial periods for river systems that discharge to the northern GOM. Key elements of this thesis are as follows:

1. Published sediment volumes for a number of shelf-margin deltas are first used to determine the mass of sediment stored in shelf-margin deltas, and published geochronological data are used to estimate deposition rates.
2. Recently developed empirical models of sediment supply (Syvitski et al., 2003; Syvitski and Milliman, 2007) are used to estimate supply rates for modern river systems and the role of glacial period boundary conditions on supply rates. Modeling incorporates

published estimates of decreases in temperature, sea-level lowering to mid-shelf or outer-shelf positions, and increased drainage area due to shelf exposure and merging of drainage basins on the shelf during sea-level fall.

3. The spectrum of possible sediment fluxes derived from the Syvitski and Milliman (2007) model are then used to ascertain changes in sediment supply necessary to account for observed sediment mass in shelf-margin deltas.

BACKGROUND

Study Area

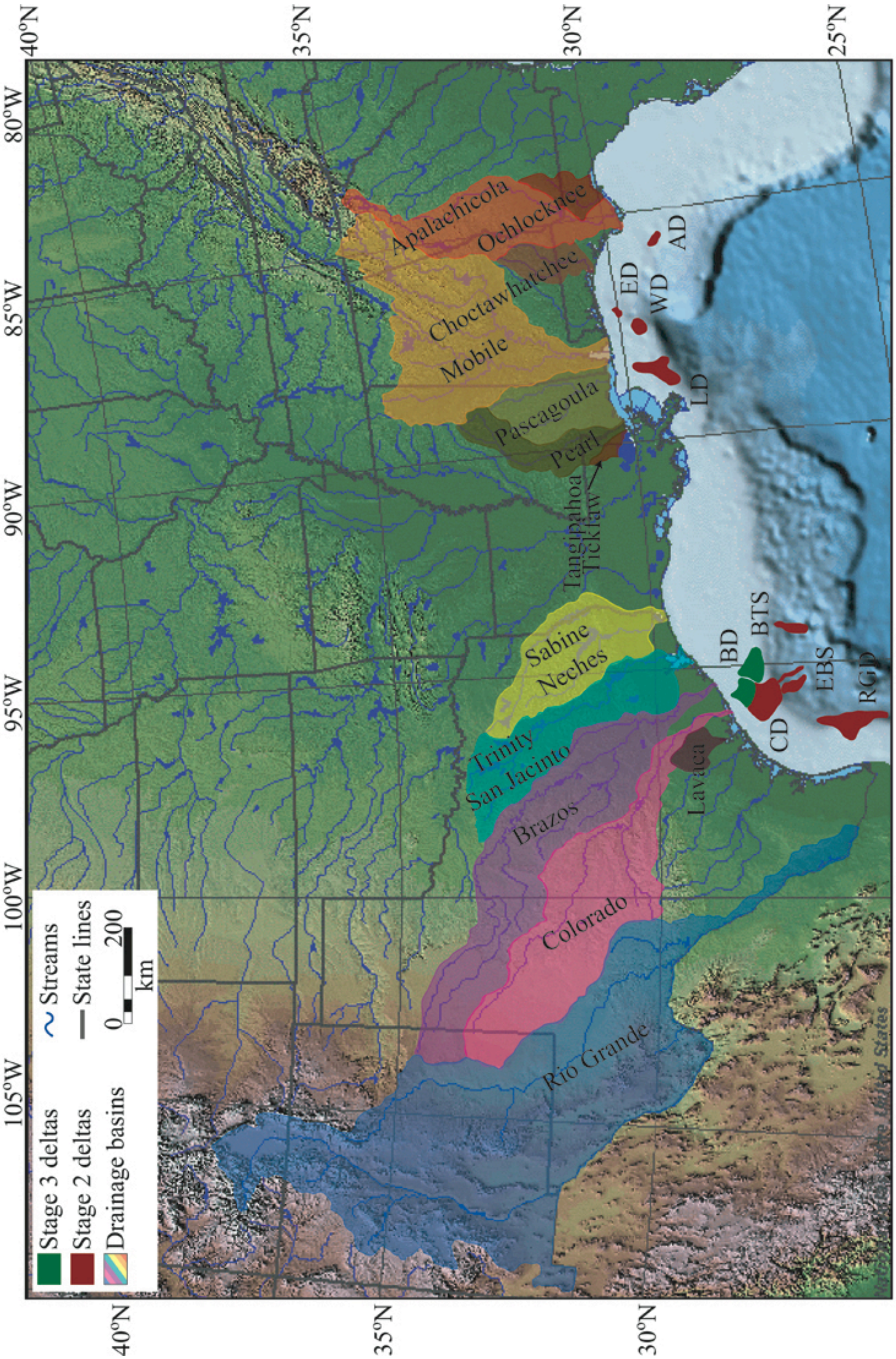
The GOM is a partially enclosed sea that covers an area of roughly 1,600,000 km^2. The northern margin of the GOM, the focus of this thesis, includes the northwestern coast of Florida, and the coasts of Alabama, Mississippi, Louisiana and Texas (Fig. 1). The northern GOM is a passive continental margin, with a low gradient coastal plain and shelf that is nourished by river systems of varying scale and complexity (Fig. 2). The following discussion provides an overview of the origin and evolution of the GOM, and the present day river systems that deliver sediment to the northern GOM margin.

Geologic Setting

Formation of the GOM began in the late Triassic with the break up of Pangea (Buffler, 1991; Salvador, 1991). Crustal extension and thinning resulted in the initial rifting of the basin, which was accompanied by the formation of oceanic crust and deposition of evaporites from the late Middle Jurassic to the early Cretaceous (Salvador, 1991). This phase was followed by a period of subsidence, sea level transgression, and shallow-marine carbonate deposition along the northern margin, which was receiving limited terrigenous clastic input from the Ouachita and Appalachian mountain belts (Martin, 1978; Buffler, 1991; Salvador, 1991). The late Jurassic transgression, at its maximum during the early Cretaceous, connected the GOM basin with the Pacific Ocean, Atlantic Ocean, as well as the Western Interior Seaway, which continued communication periodically until the onset of the Laramide Orogeny during the late Cretaceous (Salvador, 1991).

The Cenozoic history of sedimentation along the northern margin reflects the interplay of various factors including: tectonic activity in the hinterland, climate fluctuations, and sea-level

Figure 1. Location map showing shaded relief, drainage basins, deltas associated with each, and general bathymetry. The different deltas are labeled as follows: AD – Apalachicola delta (McKeown et al., 2004), ED – eastern delta (Bart and Anderson, 2004), WD – western delta (Bart and Anderson, 2004), LD – Lagniappe delta (Roberts et al., 2000), BD – Brazos delta (Fraticelli, 2004), BTS – Brazos-Trinity slope system (Beaubouef et al., 2003), CD – Colorado delta (van Heijst et al., 2001), EBS – east breaks slide (van Heijst et al., 2001), RGD – Rio Grande delta (Banfield and Anderson, 2004). Base map modified from National Atlas of the United States (2007).


Stage 3 deltas
Stage 2 deltas
Drainage basins
Streams
State lines
0
200
km
40°N
35°N
30°N
25°N
80°W
85°W
90°W
95°W
100°W
105°W
Apalachicola
Ochlocknee
Choctawhatchee
Mobile
Pascagoula
Pearl
Tangipahoa
Ticklaw
Sabine
Neches
Trinity
San Jacinto
Brazos
Colorado
Lavaca
Rio Grande
ED
WD
AD
LD
BD
BTS
EBS
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RGD

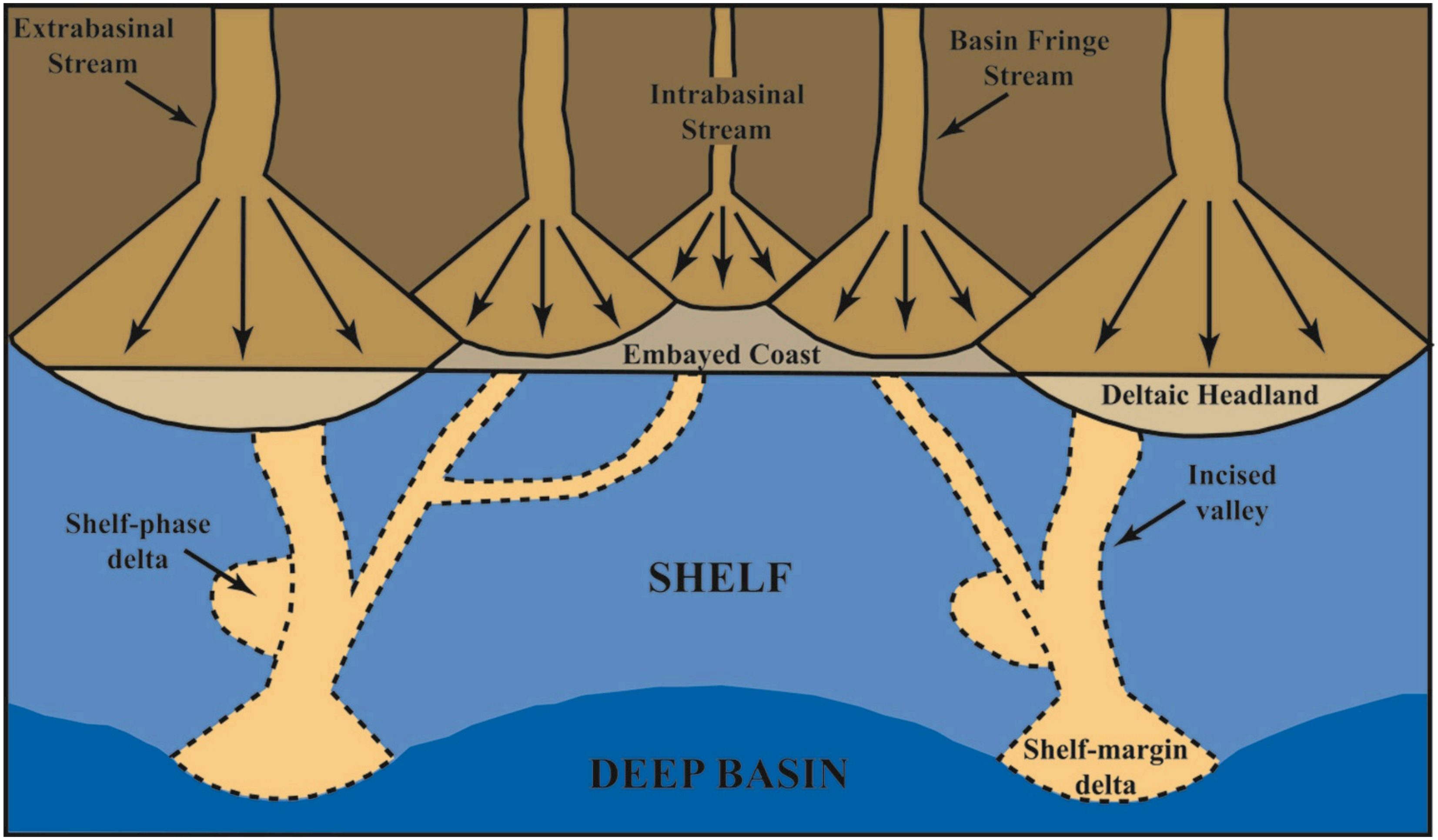


Figure 2. Model for large-scale morphology and depositional systems typical of passive continental margins. It depicts large subaerial alluvial-deltaic headlands emerging from valleys with extrabasinal source areas, embayed coasts with small alluvial-deltaic plains with basin fringe and intrabasinal source areas, incised valleys cut by rivers during sea-level fall, shelf-phase deltas as well as shelf0margin deltas. First developed by Winker (1979) for the Texas Gulf Coastal Plain and shelf and later adapted by Blum and Price (1998) using various studies from the northern GOM. Modified from Blum and Price (1998).

rise and fall (Martin, 1978; Galloway, 1981; Winker, 1982; Galloway et al., 2000; Galloway, 2005). All of these variables produced changes in sediment supply from the five principal fluvial drainages that have essentially delivered sediment to the basin margin since the Paleocene: minor catchments draining the coastal plain were, by contrast, most likely only affected by sea level and regional climate (Galloway et al., 2000; Galloway, 2005). The Rio Grande, Houston, Red, Mississippi, and Tennessee fluvial axes were the main conduits of sediment from the continental interior. Sediment delivery through these axes resulted in progradation of the shelf margin, from the Cretaceous carbonate-platform deposits to the present shelf margin, which is roughly 100 km, on average, from the present coastline. Episodic interruptions in progradation of the continental margin occurred due to salt withdrawal, mass wasting (Martin, 1978; Galloway, 1981; Winker, 1982; Galloway et al., 2000; Galloway, 2005), and/or variations due to changes in orogenic activity and sediment flux (Winker, 1982).

Present-Day River Systems and Drainage Basins

Bridge (2003) defines a drainage basin as the area that contributes water and sediment to a river system, hereafter called contributing area. Fetter (1994) states that the drainage basin consists of all the land area sloping toward a particular discharge point including those areas that do not contribute water and sediment, hereafter called noncontributing area. Taking into consideration these two definitions, it is no surprise that there are discrepancies between the values reported for the catchment of each river along the Gulf Coast with some researchers reporting the entire drainage area (e.g. Milliman and Syvitski, 1992; Syvitski and Milliman, 2007) while others (e.g. USGS, 2007) report only the area that contributes water and sediment. This thesis uses area estimates encompassing only contributing drainage.

The northern GOM is fed by river systems of varying size (Fig. 1): as noted above, most major rivers have a long history of drainage network integration and long-lived structurally focused entrances to the basin margin (Galloway, 1981; 2005). Winker (1982) and Galloway (1981) classify river systems of the region as extrabasinal (Fig. 2) if they have larger, well-integrated drainage areas that begin in higher-relief source terrains farther inland than the limits of the Cenozoic GOM basin fill, basin-fringe if they have smaller well-integrated drainage networks that mostly drain the inland margins of the basin fill, and intrabasinal if they originate on the Gulf Coastal Plain, and have small drainage areas. Blum and Price (1998) modified Winker's (1982) model, noting that extrabasinal streams are the axial streams that were likely joined by smaller systems as they traversed the shelf during periods of sea-level fall and lowstand (Fig. 2).

The Mississippi is the axial river for a continental-scale system, which drains much of the continental interior of North America, and with other ancestral rivers, is responsible for much of the sediment accumulated in the GOM basin. The Mississippi drainage basin crosses several climatic zones and includes areas that were glaciated over the last several glacial cycles. Due to the complexity of the Mississippi system, it is less suitable for the type of analysis conducted by this thesis and thus is not included here.

To the east of the Mississippi River, major extrabasinal systems include the Apalachicola and Mobile, whereas to the west, major extrabasinal systems include the Brazos, Colorado, and Rio Grande (Fig. 1). Each of these drainage basins has significant inland drainage areas, as well as well-integrated drainage networks, and represents major sources for sediment input into the depositional basin. Each extrabasinal system is bounded by smaller basin-fringe and intrabasinal systems that discharge separately to the coastal oceans today, but may have acted as tributaries to the larger systems during sea-level lowstand (Blum and Price, 1998; Blum and Aslan, 2006). Key characteristics of these river systems are summarized below, as taken from Hovius (1998), Syvitski and Milliman (2007), and the United States Geological Survey (USGS; 2007).

The Apalachicola system

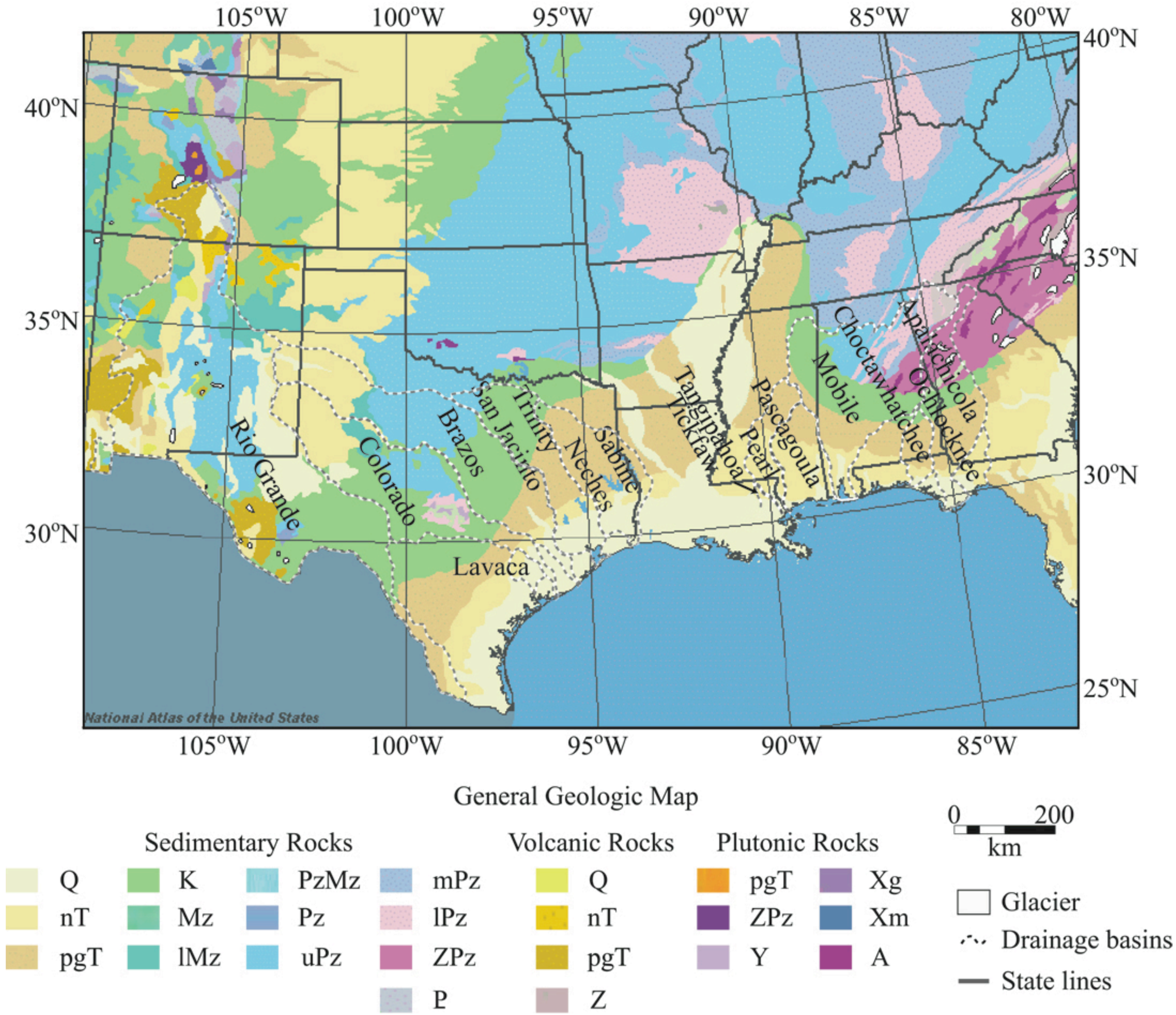
The Apalachicola (Fig. 1) drains an area of 50,000 km^2 (USGS, 2007), within the Piedmont of the southern most Appalachians physiographic province, then crosses the Gulf Coastal Plain before discharging to the GOM at the town of Apalachicola, in northwest Florida (Anderson, 2001; McKeown et al., 2004). Maximum relief in the drainage basin is 1.245 km (USGS, 2007), whereas rock types (Fig. 3) that contribute to the fluvial sediment supply include Precambrian and Paleozoic meta-sedimentary units and volcanic rocks with granitic intrusions (Press et al., 2003) within the southern Appalachians, and Cretaceous carbonates, calcareous mudstones, and sandstones as well as Cenozoic clastic sedimentary rocks within the Gulf Coastal Plain (Galloway et al., 2000; USGS, 2005).

Some rivers possibly merged with the Apalachicola during lowstands: these include the Choctawhatchee River to the west and the Ochlocknee River to the east. The modern Apalachicola River discharges to a bayhead delta, and has not yet filled its coastal plain incised valley from the last glacial period (Donoghue, 1989). Numerous dams affect sediment discharge to the river mouth, with the farthest downstream dam being Jim Woodruff Dam completed in 1952. Jim Woodruff Dam captures almost 90% of the Apalachicola drainage area and its reservoir has a storage capacity of 0.5 km^3 (US Army Corp of Engineers, 1995).

The Mobile system

The Mobile River (Fig. 1) forms from the merging of the Alabama and Tombigbee Rivers just upstream from the city of Mobile. Through this confluence the Mobile drains an area

Figure 3. Generalized geologic map of the southeastern and southcentral United States. The different units are labeled as follows: Q – Quaternary, nT – Neogene, pgT – Paleogene, K – Cretaceous, Mz – Mesozoic, lMz – Lower Mesozoic, PzMz – Paleozoic and Mesozoic, Pz – Paleozoic, uPz – upper Paleozoic, mPz – middle Paleozoic, lPz – lower Proterozoic, ZPz – Late Proterozoic, P – Proterozoic, Z – late Proterozoic, Y – middle Proterozoic, Xg – early Proterozoic granites, Xm – late Proterozoic mafics, and A – Archean. Modified from National Atlas of the United States (2007).


105°W
100°W
95°W
90°W
85°W
80°W
40°N
35°N
30°N
25°N
Rio Grande
Colorado
Brazos
San Jacinto
Trinity
Neches
Sabine
Lavaca
Tangipahoa
Tickfaw
Pearl
Pascagoula
Mobile
Choctawhatchee
Ochlocknee
Apalachicola
National Atlas of the United States
General Geologic Map
Sedimentary Rocks
Q
nT
pgT
K
Mz
lMz
PzMz
Pz
uPz
mPz
lPz
ZPz
P
Volcanic Rocks
Q
nT
pgT
Z
Plutonic Rocks
pgT
ZPz
Y
Xg
Xm
A
0
200
km
Glacier
Drainage basins
State lines

of 113,500 km^2, within the Valley and Ridge province, Piedmont, and Cumberland Plateau of the Appalachian fold belt physiographic region (USGS, 2007), before traveling across the Gulf Coastal Plain to discharge at Mobile, in south central Alabama. Maximum relief in the drainage basin is 1.188 km (Hovius, 1998), whereas rock types (Fig. 3) that contribute to the fluvial sediment supply include Precambrian and Paleozoic meta-sedimentary and volcanic rocks with granitic intrusions from the southern Appalachians (Johnson et al., 2002; Press et al., 2003). Cretaceous carbonates, calcareous mudstones, and sandstones as well as Cenozoic clastic sedimentary rocks and Paleogene clastics are derived from within the Gulf Coastal Plain (Galloway et al., 2000; Johnson et al., 2002; USGS, 2005).

Some rivers possibly merged with the Mobile during lowstands: these include the Escambia, Blackwater, and Yellow Rivers in the east as well as the Pascagoula, Pearl, Tangipahoa, Tickfaw, and Amite Rivers to the west. The modern Mobile River discharges to a bayhead delta and has not yet filled its coastal plain incised valley from the last glacial period. Numerous dams affect sediment discharge to the river mouth, with the farthest downstream dam being Claiborne Dam completed in 1969. Claiborne Dam captures approximately 50% of the Mobile drainage area and its reservoir has a storage capacity of 0.12 km^3 (US Army Corp of Engineers, 1995).

The Brazos system

The Brazos (Fig. 1) drains an area of 117,400 km^2 (USGS, 2007), within the Southern High Plains, North Central Plains and Grand Prairie physiographic provinces, then transects the Gulf Coastal Plain before discharging to the GOM approximately 15 km southwest of the city of Freeport, in southeast Texas (BEG, 1992; 1996). Maximum relief in the drainage basin is 1.441 km (USGS, 2007), whereas rock types (Fig. 3) that contribute to the fluvial sediment supply include Permian carbonates and siliciclastics within the North Central Plains, Triassic siliciclastics, Cretaceous carbonates and sandstones within the Grand Prairie province. Cretaceous carbonates and calcareous mudstones as well as Cenozoic clastic sedimentary rocks are derived from within the Gulf Coastal Plain (Galloway, 1981; BEG, 1992; Winker, 1982; BEG, 1996; USGS, 2005).

Some rivers possible merged with the Brazos during lowstands: these include the San Jacinto, Trinity, Neches, and Sabine Rivers (Fig. 1), all located to the east of the modern drainage outlet. The modern Brazos River has filled its coastal plain incised valley from the last glacial period and discharges to a wave-dominated delta located along the northwestern GOM margin. Numerous dams affect sediment discharge to the river mouth, with the farthest downstream dam being Whitney Dam completed in 1951. Whitney Dam captures slightly less than 40% of the Brazos drainage area and its reservoir has a storage capacity of 0.77 km^3 (US Army Corp of Engineers, 1995).

The Colorado system

The Colorado (Fig. 1) drains an area of 109,400 km^2 (USGS, 2007), within the Southern High Plains and Edwards Plateau physiographic provinces, then crosses the Balcones Escarpment before it transverses the Gulf Coastal Plain and discharges to the GOM less than 10 km south of the city of Matagorda, in southeast Texas (BEG, 1992; Blum, 1993; BEG, 1996). Maximum relief in the drainage basin is 1.344 km (Syvitski and Milliman, 2007), whereas rock types (Fig.3) that contribute the majority of the fluvial sediment supply include Triassic and Permian siliciclastics, Permian carbonates, Pennsylvanian carbonates and clastics, lower Paleozoic sedimentary rocks, and Cretaceous carbonates and sandstones within the Edwards Plateau. A minor contribution of sediment of Cretaceous carbonates and calcareous marine mudstones, and Cenozoic clastic sedimentary rocks come from the Gulf Coastal Plain (Galloway, 1981; Winker, 1982; BEG, 1992; Blum, 1993; BEG, 1996; USGS, 2005).

Some rivers possibly merged with the Colorado during lowstands: these include the Lavaca River to the south and the San Bernard River to the east. The modern Colorado River has filled its coastal plain incised valley formed during the last glacial period and discharges to a wave-dominated delta along the northwestern GOM margin. Numerous dams affect sediment discharge to the river mouth, with the farthest downstream dam being the Marshall Ford Dam completed in 1942. Marshall Ford Dam captures nearly 90% of the Colorado drainage area and its reservoir has a storage capacity of 1.7 km^3 (US Army Corp of Engineers, 1995).

The Rio Grande system

The Rio Grande (Fig. 1) drains an area of 471,895 km^2 (USGS, 2007), within the Southern Rocky Mountains, San Luis Valley, Rio Grande Rift, Basin and Range, Southern High Plains, and Edward Plateau physiographic provinces, then transects the Gulf Coastal Plain before discharging to the GOM 70 km east of the city of Brownsville, in south Texas (BEG, 1992; 1996). Maximum relief in the drainage basin is 4.188 km (USGS, 2007). Rock types (Fig. 3) that contribute to the fluvial sediment supply include Tertiary and Cretaceous volcanics within the Southern Rocky Mountains and San Luis Valley. Farther south, Tertiary and Quaternary volcanics as well as Quaternary clastics are derived from within the Rio Grande Rift Valley. From the Basin and Range Province, Quaternary clastics and volcanics, Tertiary volcanics, Cretaceous carbonates, and Permian carbonates and siliciclastics are contributed while Triassic and Permian siliciclastics, Permian carbonates, Pennsylvanian carbonates and clastics, lower Paleozoic sedimentary rocks, and Cretaceous carbonates and sandstones come from the Edwards Plateau. Cretaceous carbonates and calcareous marine mudstones, and Cenozoic clastic sedimentary rocks are supplied from within the Gulf Coastal Plain (Galloway, 1981; BEG, 1992; Winker, 1982; BEG, 1996; USGS, 2005).

There are no nearby drainage basins that are candidates for merging with the Rio Grande during lowstands. The modern Rio Grande has filled its coastal plain incised valley formed during the last glacial period and discharges to a delta along the western margin of the GOM. Numerous dams affect sediment discharge to the river mouth, with the farthest downstream being International Falcon Dam completed in 1953. International Falcon Dam captures over 85% of the Rio Grande drainage area and its reservoir has a storage capacity of 3.3 km^3 (US Army Corp of Engineers, 1995).

The Shelf and Shelf Margin

Cenozoic progradation of the continental margin has constructed a substantial shelf around the GOM (Fig. 1). Carbonate sediment is dominant off the coast and shelf of the Florida peninsula, reflecting the lack of clastic input in that area from major river systems; whereas, terrigenous clastics underlie the shelf from NW Florida to Mexico (Holcombe et al., 2002). The ramp-like continental shelf from northwest Florida to Alabama narrows from roughly 150 km off west Florida to about 30 km near the Florida-Alabama border. The shelf widens for a short distance before narrowing again due to progradation of the modern Mississippi delta to the shelf margin. The Texas-Louisiana shelf, with an average width of 100 km, is characterized by a broad platform with a low gradient, roughly 0.5 m/km, off the coast of west Louisiana and east Texas, and a steep (1.2-2.75 m/km) gradient and narrow width off the central and south Texas coast (Anderson et al., 2004).

Broad clastic shelves and shelf margins like that of the GOM are a result of the transit of fluvial-deltaic systems across the shelf during sea-level falls and lowstands, as well as waves that sculpt the surface (Muto and Steel, 2004; Porebski and Steel, 2004; 2006). The GOM figured prominently in the development of this concept, based on the early definition of shelf-phase and shelf-margin deltas (Fig. 2) in offshore Texas by Winker (1982) and Suter and Berryhill (1986). For the last glacial period, each of the river systems discussed herein is thought to have cut through highstand shorelines (Fig. 4), transited the shelf, and constructed deltas on the shelf and at the shelf margin (Anderson and Fillon, 2004, Anderson, 2005; Blum and Aslan, 2006). The links between specific river systems onshore and specific deltas in the offshore, however, are not always well established.

The Apalachicola delta (Fig. 1) is believed to be deposited by the Apalachicola River from 75,000-25,000 years BP and is located southwest of the modern Apalachicola depocenter (Anderson, 2001; McKeown et al., 2004). The eastern delta and western delta (Fig. 1; Sager et al., 1999) are thought to be products of the Mobile River deposited sometime between 75,000 years BP and 16,000 years BP (Sager et al., 1999; Correa-Lafayette, 2001; Bart and Anderson, 2004; Bartek et al., 2004), whereas the Lagniappe delta, roughly 70 km to the west (Fig. 1), appears to be also connected to the Mobile River with an age, based on accelerated mass spectrometry (AMS) ^{14}C dates from cores, of 23,000-19,000 years BP (Kindinger, 1988; 1989; Kindinger et al., 1989; Sydow, 1992; Sydow et al., 1992; Sydow and Roberts, 1994; Roberts et al., 2004). The Brazos delta (Fig. 1) is linked to the Brazos River between 44,400 years BP and 33,720 years BP (Anderson, 2005; Fraticelli, 2004; Fraticelli and Anderson, 2003), after which the Brazos joined the Trinity and Sabine Rivers (Abdulah, 1995; Abdulah et al., 2004; Fraticelli, 2004; Anderson, 2005) to fill a series of intraslope minibasins termed the Brazos-Trinity slope system (Fig. 1) during an expansive time frame of 100,000 years (115,000-15,000 years BP; Badalini et al., 2000; Beaubouef and Friedmann, 2000; Beaubouef et al., 2003; Mallarino et al.,

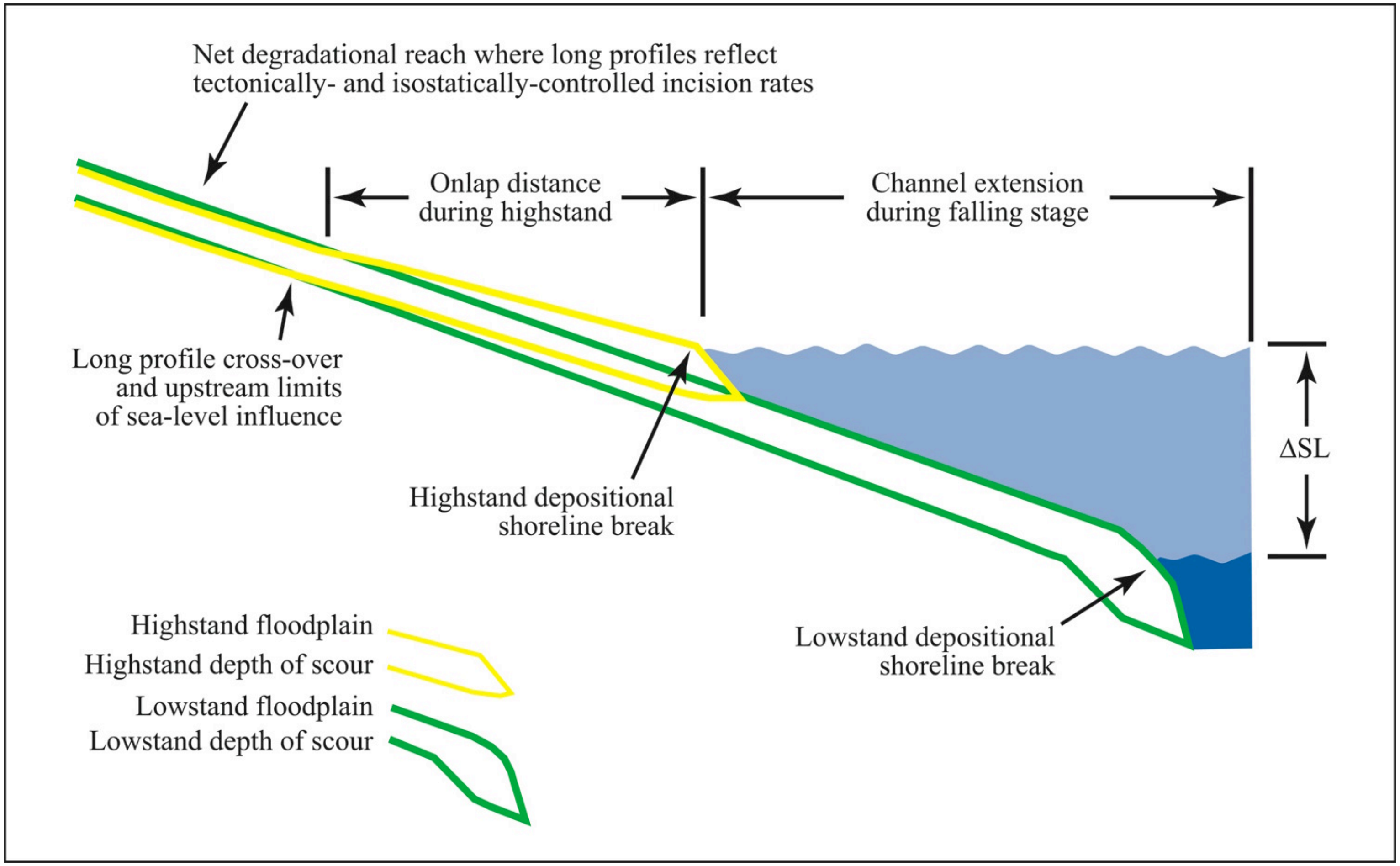


Figure 4. Definition sketch for fluvial response to sea-level change (ΔSL) along a continental margin with a distinct highstand depositional shoreline break. Diagram illustrates concepts of channel extension during sea-level fall and lowstand, versus upstream limits of onlap during sea-level rise and highstand. Upstream from this cross-over point, channel long profiles are independent of high-frequency sea-level changes, and reflect long-term tectonically controlled base level of erosion. Modified from Blum and Price (1998).

2006). The Colorado River deposited two deltas (Fig. 1) trending southeast of its present point of discharge (Abdulah, 1995; Snow, 1998; van Heijst et al., 2001; Abdulah et al., 2004). The delta on the middle shelf developed from 40,000-23,000 years BP as sea level fell (van Heijst et al., 2001), whereas the other located nearer the shelf edge was constructed from 23,000-11,500 years BP (van Heijst et al., 2001). The Rio Grande is associated with sediment on the outer shelf and upper slope (Fig. 1), east of its present position, deposited from 19,000 years BP to 11,000 years BP (Banfield, 1998; Banfield and Anderson, 2004).

Subsidence

Overall the northern GOM continental margin is tectonically stable with subsidence rates that vary depending on the amount and rate of sediment accumulation, salt remobilization, shale diapirism, and growth faulting (Martin, 1978; Abdulah et al., 2004). Off the coast of west Florida, the lack of an abrupt shelf break, a relict of the seaway through north Florida connecting the Atlantic to the Gulf that persisted until recently (Galloway et al., 2000), has not allowed the accumulation of large amounts of sediment. This factor, coupled with no significant subsurface salt, has resulted in low subsidence rates. Further to the west in the region of the modern Mississippi delta, episodic salt evacuation caused by sediment loading (e.g. Woodbury et al., 1973; Morton and Suter, 1996), Holocene sediment compaction (e.g. Fisk, 1954; Meckel et al., 2006), fluid withdrawal (e.g. Morton et al., 2002), and movement along subsurface faults (e.g. Dokka, 2006) have all been connected to high subsidence rates. The Texas coast and shelf have experienced low (0.05-1 mm/yr) long-term average subsidence rates throughout the Quaternary (Abdulah et al., 2004; Anderson et al., 2004); whereas, low to high (1-22 mm/yr) subsidence rates have been measured over decadal time scales (Paine, 1993).

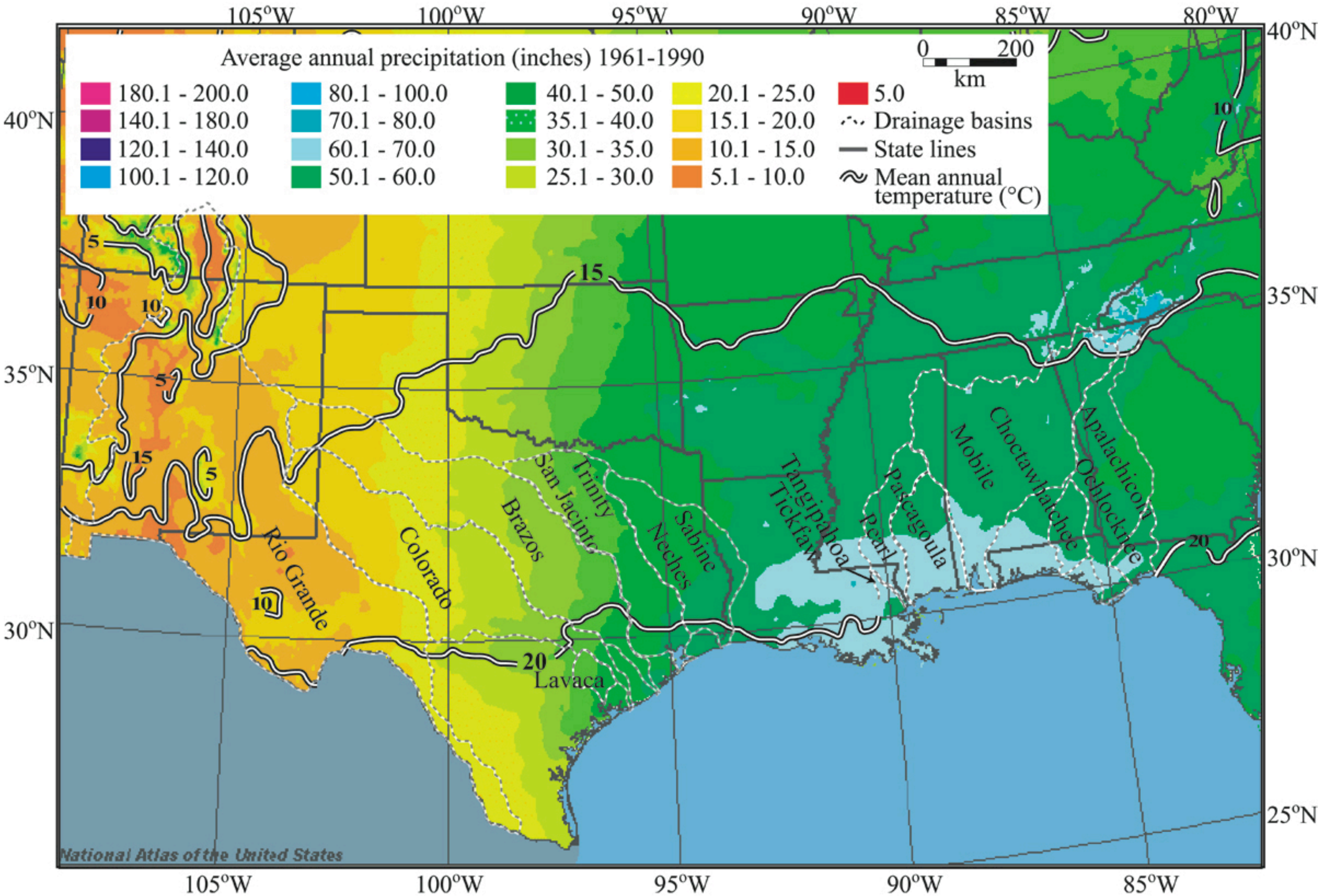
Climate and Hydrology

Modern climate across the northern Gulf Coast varies from subtropical humid in the east to subtropical semiarid in the west, with northern parts of the various drainage basins having more temperate climatic regimes. At a very general level, mean annual precipitation along the Gulf Coast decreases from east to west, although the highest annual values are found within the coastal region of Louisiana, Mississippi, and Alabama (Fig. 5; National Oceanic and Atmospheric Administration, 1995; National Atlas of the United States, 2007).

The northeastern margin of the GOM includes the Apalachicola and Mobile River basins, described above, plus a number of smaller basin-fringe and intrabasinal systems. At a more detailed level the climate is mainly humid temperate in the north to humid subtropical in the south with mean annual temperatures of 18°C and 17°C (Fig. 5), respectively. The precipitation maxima occur in the summer months for both basins (National Oceanic and Atmospheric Administration, 1995; National Atlas of the United States, 2007).

The northwestern margin of the GOM includes the Brazos River, Colorado River, and Rio Grande drainage basins, plus a number of smaller basin-fringe and intrabasinal systems. Both the Brazos and Colorado Rivers flow through several climate zones, from semiarid temperate in the northwest to humid subtropical in the southeast, with mean annual temperatures of 19°C and 18°C respectively (Fig. 5; Hovius, 1998). Farther west, the Rio Grande experiences arid conditions through most of the drainage area, and a mean annual temperature of roughly 15°C (Fig. 5), but is semiarid subtropical in the lower Rio Grande Valley. Precipitation over this region is as variable as the climate, but again decreases from east to west, with the lowest mean annual precipitation occurring within the Rio Grande catchment (Fig. 5).

Figure 5. Map of average annual precipitation in inches and mean annual temperature in degrees Celsius. Modified from National Atlas of the United States (2007).



Average annual precipitation (inches) 1961-1990
180.1 - 200.0
140.1 - 180.0
120.1 - 140.0
100.1 - 120.0
80.1 - 100.0
70.1 - 80.0
60.1 - 70.0
50.1 - 60.0
40.1 - 50.0
35.1 - 40.0
30.1 - 35.0
25.1 - 30.0
20.1 - 25.0
15.1 - 20.0
10.1 - 15.0
5.1 - 10.0
5.0
Drainage basins
State lines
Mean annual temperature (°C)
0
200
km
Rio Grande
Colorado
Brazos
San Jacinto
Trinity
Lavaca
Neches
Sabine
Tangipahoa
Tickfaw
Pearl
Pascagoula
Mobile
Choctawhatchee
Ochlockne
Apalachicola
National Atlas of the United States
105°W
100°W
95°W
90°W
85°W
80°W
40°N
35°N
30°N
25°N

Discharge at a river mouth is largely a function of drainage-basin size and the climatic parameters of temperature and precipitation, though geologic factors also play a role. Modern discharge values are strongly influenced by numerous dams that have been constructed on all rivers of interest for this thesis, and various compilations show different values depending on the time series of data used (contrast Hovius, 1998 with Syvitski and Milliman, 2007). However, in general, specific discharge (runoff per unit drainage area) corresponds to the precipitation patterns described above (Fig. 5), with the Mobile and Apalachicola River having the highest values, and Rio Grande having the lowest values. Hovius (1998) and Syvitski and Milliman (2007) indicate the Mobile and Apalachicola Rivers have the highest mean annual discharge due to a humid subtropical climate and the highest annual precipitation in the study area (Hovius, 1998), whereas the Colorado River has the lowest, due to its semiarid headwaters. Mean annual discharge on Rio Grande is smaller than that of all other systems, except the Colorado, in spite of the significantly larger drainage area, due to the semiarid climate.

Controls on Sediment Supply

Sediment supplied by river systems to the coastal ocean is a function of drainage area, relief, basin lithology, and climate (Milliman and Syvitski, 1992; Hovius, 1998; Syvitski et al., 2003; Syvitski and Milliman, 2007). Many researchers (e.g. Milliman and Syvitski, 1992; Summerfield and Hulton, 1994; Mulder and Syvitski, 1996; Hovius, 1998; Leeder et al., 1998; Harrison, 2000) have established that the first order controls are area, relief, sediment discharge, and lithology all of which ultimately determined by tectonic activity (Ahnert, 1960; Pinet and Souriau, 1988; Milliman and Syvitski, 1992). Drainage area is the result of the tectonic organization of hinterland source terrains, whereas relief is a result of tectonic, isostatic, and erosional processes. Increases in drainage area and relief, in general, produce higher sediment supplies (Ahnert, 1970; Milliman and Meade, 1983; Milliman and Syvitski, 1992), albeit for different reasons.

A large drainage area has a greater number of tributaries contributing to the primary trunk stream thus increasing the amount of sediment supplied to the system. Relief, on the other hand, is the primary control on rates of erosion or sediment production (Ahnert, 1970; Pinet and Souriau, 1988; Milliman and Syvitski, 1992), such that mechanical erosion rates increase linearly with basin elevation (Pinet and Souriau, 1988). Lithology is another key factor in determining the rate of soil and rock erosion: soft rocks (sedimentary rocks) are easily erodible, whereas hard rocks (igneous and metamorphic rocks) are more difficult to break down. The response time of each system between the transference of sediment from the uplands to the basin controls the rate at which sediment is brought to the basin (Leeder et al., 1998; Castelltort and Van Den Driessche, 2003; Syvitski, 2003).

Climate parameters, such as temperature and precipitation, are secondary controls on sediment discharge (Syvitski et al., 2003; Syvitski and Milliman, 2007): they affect rates of weathering, the location, amount and phase of precipitation, the type and density of vegetation, and the amount of runoff or water discharge (Langbein and Schumm, 1958; Wilson, 1973). With all other factors equal, the lower midlatitudes and tropics, in general, would experience an increase in sediment supply with increasing temperature or increasing water discharge, while the upper midlatitudes and subpolar regions would experience an increase in sediment supply with increasing glaciation. Holding temperature constant, sediment supply would also increase with increasing precipitation and runoff (Meybeck et al., 2003; Syvitski et al., 2003; Syvitski and Milliman, 2007).

Sediment supply can be measured in a number of different ways. Sediment yield is a measure of the mass of sediment per unit area per unit time, whereas sediment load or discharge is a measure of mass per unit time. Sediment discharge is a function of drainage area while sediment yield is a function of tectonic setting and rates of rock uplift. In fact, sediment yields are highest in contractional orogens with yields up to 2 orders of magnitude greater than low-relief cratons where the majority of large drainage areas are located (Hovius, 1998). Milliman and Syvitski (1992) established that the majority of sediment delivered to the oceans is derived from the many short, steep rivers within active orogens. This is a direct function of sediment yield decreasing as drainage area increases due to an increase in sediment storage (Milliman and Meade, 1983; Milliman and Syvitski, 1992). This illustrates that, although large systems almost always discharge more sediment than smaller systems, it is the small, steep systems that account for the majority of the total sediment produced and transferred to the coastal ocean (Milliman and Syvitski, 1992).

In previous studies, comparisons between sedimentation rates from shelf and shelf-margin deltas were compared with sediment supply measured from modern streams (e.g. Banfield, 1998; Snow, 1998; Correa-Lafayette, 2001; Abdulah et al., 2004; Banfield and Anderson, 2004; McKeown et al., 2004). This approach produces a potential source of error, since the sediment discharge values used were strongly biased towards the period after dams were constructed on many streams, and therefore significantly underestimates natural pre-dam sediment yield.

Hovius (1998) and Syvitski and Milliman (2007) compile data on sediment yields and sediment discharge for a large number of river systems, including the GOM systems examined here. Both values are measured at varying distances upstream from the mouth of the river, and measurements do not include either sediment transported as bedload, or sediment stored in the floodplain between the point of measurement and the river mouth. For this thesis, however, values measured and reported in these compilations are assumed to represent sediment supply to the fluvial-deltaic depositional system (Fig. 4), and the bedload component is assumed to represent 10% (Judson and Ritter, 1964). Pre-dam sediment supply data are sparse and, for example, the state of Texas contains 6,500 dams, which is more than any state within the United States (United States Army Corps of Engineers, 1995).

LeBlanc and Hodgson (1959) and Kanes (1970) compiled early data on sediment loads for major streams in Texas, which show the Brazos River delivered an average of 31 Mt/yr (for the period 1924-1954) followed by the Colorado River at ~13 Mt/yr (1931-1945). Some of these data were used in Hovius' (1998) global compilation, whereas other workers used more complete data sets that include the post-dam period, which underestimate natural sediment loads for some of these systems (e.g. Milliman and Syvitski, 1992; Syvitski and Milliman, 2007). As for the Mobile River and Apalachicola River, no pre-dam sediment loads are reported. Judson and Ritter (1964) report an average annual suspended load of 3.1 Mt/yr, measured at Claiborne, Alabama, for the Mobile River between 1952 and 1960. However, in 1955 Demopolis Lock and Dam was completed capturing roughly 35% of the Mobile drainage (United States Army Corps of Engineers, 1995). The average load carried by the Apalachicola River, on the other hand, has been calculated using the volume contained within the delta (1.2 MT/yr; McKeown et al., 2004) and sedimentation rates within Apalachicola Bay (1.5 MT/yr; Isphording, 1986). These measurements provide the only available basis for comparison between measured and modeled values of discharge and sediment load.

In the absence of reliable pre-dam data, an alternative approach would be to estimate modern pre-dam sediment supply using empirical models. Syvitski et al. (2003) present a first-generation empirical model for estimation of the long-term flux of sediment from river basins to the coastal ocean, then illustrate its application to the Eel River of coastal California, and the Fly River, Papua New Guinea. This model, referred to as ART (area, relief, and temperature), enables the user to calculate average discharge using:

$$Q = \alpha_1 A^{\alpha_2} \tag{1}$$

where: $Q =$ average discharge (m^3/s)

$A =$ drainage basin area (km^2)

$\alpha =$ regression coefficients

Average discharge is then added as an input to determine long-term sediment flux using:

$$Qs = \alpha_6 Q^{\alpha_7} R^{\alpha_8} e^{kT} \tag{2}$$

where: $Qs =$ long term sediment load (kg/s)

$R =$ maximum relief (m)

$k =$ regression coefficient

$T =$ basin-average temperature (°C)

This model explains 57% of the between-river variance in a dataset that includes 340 rivers that discharge to the coastal ocean from different geologic and climatic settings around the world. The ART model is a predictive tool and by no means free from error, but more importantly the source of the uncertainty, and therefore error can be identified. This model provides a maximum sediment flux to the coastal ocean, because it is based on gauging stations that are typically somewhat inland from river mouths and do not account for more seaward storage sites, such as deltaic plains, estuaries, and tidal flats. For the purposes of this thesis, then, the value predicted might be considered to be representative of the sediment discharge to updip limits of a delta plain (Fig. 4).

Syvitski and Milliman (2007) present a second-generation global predictor of sediment flux, the BQART model, from an upgraded database of 488 rivers that represents 66% of the sediment delivered to the coastal ocean as well as 63% of the land surface. The new model builds upon the ART model but (a) refines the accounting for geologic factors, by adding lithology to relief and drainage area, (b) accounts for glacial cover and human activities, and (c) retains the significance of temperature (Syvitski and Milliman, 2007).

Like the ART model, the BQART model predicts the long-term flux of sediment delivered by rivers at or near sea level without taking into account the variability in sediment supply rate, sediment stored within the system, or sediment derived from erosion within the basin (Syvitski and Milliman, 2007). The BQART model estimates sediment discharge with:

$$Qs = \omega B A^{0.5} Q^{0.31} R T \qquad \textit{for } T \geq 2℃ \qquad (3)$$

and:

$$Qs = 2\omega B A^{0.5} Q^{0.31} R T \qquad \textit{for } T < 2℃ \qquad (4)$$

where:

$$Q = 0.075 A^{0.8} \ (\mathrm{km^3/yr}) \qquad (5)$$

$\omega =$ 0.02 *for units of kg/s*

or: $\omega =$ 0.0006 *for units of MT/yr*

$$B = IL(1 - T_E)E_h \qquad (6)$$

$I =$ glacier factor

$L =$ lithology factor

$T_E =$ trapping efficiency

$E_h =$ human-influenced soil erosion

$R =$ maximum relief (km)

The BQART model assumes three things: (1) basin area and relief are proxies for tectonic activity; (2) discharge and basin area are partly independent; and (3) basin-averaged temperature is a substitute for climate (Syvitski and Milliman, 2007). Within the database of 488 drainage basins, differences in geologic factors of drainage area, relief and lithology explain 63% of the variance, whereas basin-averaged temperature explains 10% of the between-river variance, discharge explains an additional 3%, and anthropogenic factors account for the remaining variance. The addition of variables to account for ice cover, lithology, and human influence was fundamental to improving the older ART model, such that BQART explains 96% of the variance in the 488-river dataset, and shows no systematic over- or underprediction over 6 orders of magnitude in observational values. Ice cover is not an issue for the river systems studied here, but the other variables are further discussed below.

Lithology

The addition of a lithology factor to BQART is an improvement in estimating modern sediment supply and accounts for 8% of the variance. The lithology factor depends on the basin-averaged erodibility of rocks and sediment, and varies considerably within large basins, thus increasing the uncertainty. It ranges from 0.5 to 3, producing a possible 6-fold variation in sediment discharge due to this single factor.

Anthropogenic Factors

Modern rivers are heavily influenced by human activities, which on a global scale, Syvitski et al. (2005) show that human activities have, on the one hand, increased rates of erosion due to cultivation and other practices, and on the other hand, trapped sediment behind dams. The net effect is to have decreased the delivery of sediment to the coastal oceans. Within the BQART model, two factors are used to account for anthropogenic influences. The first is the anthropogenic factor (E_h), defined as a measure of human disturbance of the landscape. This factor takes into account the competing forces of urbanization, deforestation, agricultural practices, and mining activities. Syvitski and Milliman (2007) used studies that found socioeconomic conditions, land use practices, and population density to be the causes of changes in global sediment yield to define the anthropogenic factor of each basin, which ranges from 0.3 to values greater than 2 based on the population density and gross national product per capital. They recognized the anthropogenic factor may be calculated a number of ways, and that it is dependent on other factors such as the response time of the landscape after initial perturbation, and natural events that mask the influence of humans. The BQART model, however, is not able to quantify these variations in anthropogenic affects.

The second anthropogenic factor deals with the efficiency of sediment entrapment within man-made reservoirs. Trapping efficiency is calculated using the storage capacity of a reservoir and the discharge at the head of the reservoir, which essentially compares the flux of sediment into the reservoir and the flux of sediment out of the reservoir (Syvitski and Milliman, 2007). Reservoirs are able to trap a significant amount of sediment carried by a river thus increasing the erosive power downstream of the dam (Vörösmarty et al., 2003). The BQART model is not able to predict the amount of sediment eroded in response to this change in fluvial dynamics nor does it account for the decrease in storage capacity caused by sediment accumulation since the opening of the structure (Syvitski and Milliman, 2007). The value of one minus the basin-averaged trapping efficiency can vary between 0.1 and 1. The former value applies to rivers heavily controlled while the latter value applies to rivers without large dams, as well as pre-dammed rivers.

Sediment Dispersal Systems

The BQART database consists of 488 rivers with gauging stations at some distance upstream from the actual river mouth. As previously mentioned the BQART model does not take into account sediment retention within floodplains, deltas, and estuaries that are farther downstream than the gauging stations (Syvitski and Milliman, 2007). In some cases the last gauging station for a river is located a significant distance inland. As one extreme example, the Obidos gauging station on the Amazon River is 700 km upstream from the mouth (Kuehl et al., 1986), and does not necessarily predict the precise amount of sediment discharged to the ocean.

In order to better estimate the amount of sediment sequestered in deltas at any stage of sea level, sediment dispersal systems active within the delta plain and the receiving basins must be understood. Sediment budgets have been estimated for a number of systems all over the world (e.g., Wright and Coleman, 1974; McKee et al., 1983; Bornhold et al., 1986; Kuehl et al., 1986; Alexander et al., 1991; Kesel et al., 1992; Nittrouer et al., 1995; Wright and Nittrouer, 1995; Kuehl et al., 1997; Kineke et al., 2000; Shi et al., 2003; Crockett and Nittrouer, 2004; Mullenbach et al., 2004; Gerbhardt et al., 2005; Huh et al., 2006) to determine the partitioning of sediment as it is routed through the system. Sediment budgets, averaged over decades and centuries, can measure rates of deposition, and calculate the amount of sediment sequestered on adjacent shelves and in offshore regions fairly accurately (e.g., Bornhold et al., 1986; Mullenbach et al., 2004; Gebhardt et al., 2005), but a significant portion has always remained unaccounted for. Deltaic deposits, lying between the gauging station and the locations of measurements in the nearshore and shelf, are thought to have trapped some of this missing sediment (e.g., Kuehl et al., 1986; Nittrouer et al., 1995), and more distal areas on and off the shelf are thought to sequester sediment as well.

It is worthy to note that sediment deposition and transportation are dependent on both temporal and spatial factors (Wright and Nittrouer, 1995). Sea-level changes over the last 130,000 years have affected sediment supply to the coast and dispersal mechanisms within ocean basins by changing ocean circulation (Rahmstorf, 2002), shoreline position (Lambeck et al., 2002), as well as drainage areas of contributing basins (e.g., Abdulah, 1995; Antoine et al., 2003). Climate changes, on the other hand, over this same time frame have altered storm pathways and intensities (e.g., Tudhope et al., 2001; Kukla et al., 2002; Peltier and Solheim, 2004) thus affecting overall precipitation patterns and wave dynamics. The following paragraphs give a brief overview of general concepts established in studies of sediment budgets for the Amazon (Kuehl et al., 1986; Allison et al., 1995; Nittrouer et al., 1995), Changjiang (McKee et al., 1983; Wright and Nittrouer, 1995), Eel (Alexander and Simoneau, 1999; Nittrouer, 1999; Sommerfield and Nittrouer, 1999; Crockett and Nittrouer, 2004; Mullenbach et al., 2004), Ganges-Brahmaputra (Kuehl et al., 1997), Huanghe (Bornhold et al., 1986; Alexander et al., 1991; Shi et al., 2003), Mississippi (Wright and Coleman, 1974; Kesel et al., 1992), Ob, and Yenisei (Gebhardt et al., 2005) Rivers.

Given that most gauging stations in the Syvitski and Milliman (2007) database are in the lower reaches of river systems, the first sink within most present-day depositional basins would be the highstand delta plain. The amount of sediment stored within the delta depends on sediment and water discharge, sediment grain size, channel transport capacity, and degree of mouth protrusion into the basin (Wright and Coleman, 1974; Shi et al., 2003). Shi et al. (2003) determined that the amount of sediment discharged at the mouth of the Huanghe is the main factor in determining the percentage of sediment deposited behind the delta front. Over the period from 1965 to 1974, prior to the channel avulsing to the south, ~73% of the incoming sediment was sequestered in the Diaokouhe lobe of the Huanghe (Shi et al., 2003). Values for the Ganges-Brahmaputra delta plain are also very high, estimated to be as high as 80% (Kuehl et al., 1997), whereas 10%, 33%, and 40% of total load are deposited in the delta plains of the Mississippi River, Amazon River, and Changjiang, respectively (Kesel et al., 1992; Nittrouer et al., 1995; Wright and Nittrouer, 1995).

Not all rivers discharge to the open ocean, some discharge to bayhead deltas located at the head of estuaries. In fact today most rivers delivering sediment to the coasts of the United States discharge to estuaries. Examples of rivers emptying into estuaries include the Ob and Yenisei River, which drain the northern coast of Siberia: roughly 90% of the incoming sediment is buried within these estuaries (Gebhardt et al., 2005), which greatly reduces the amount of sediment delivered to the open ocean.

Once river born sediment reaches ocean basins, sediment dispersal patterns vary spatially and temporally depending on river outflow characteristics, density contrasts between river outflow and ambient water within the basin, water depth, shelf slope, tidal range, tidal influence within the lower river channel, wave energy, and current energy (Wright and Coleman, 1974; Wright, 1977; Nittrouer and Wright, 1995; Wright and Nittrouer, 1995).

Based on sediment budget studies, averaged over a range of time scales, the continental shelf sequesters the majority of sediment delivered to the coastal ocean each year, with rates of deposition decreasing in the seaward direction (Wright and Nittrouer, 1995). Most rivers deliver sediment to the ocean via plumes that decelerate due to interaction with basin waters, which causes initial deposition of sediment load (Wright and Coleman, 1974; Wright and Nittrouer, 1995). Estimates of the percentage of sediment, both coarse and fine, that is initially deposited on the inner continental shelf (<40 m water depth) range from 25% to 90% (Wright and

Coleman, 1974; Bornhold et al., 1986; Kuehl et al., 1986; Alexander et al., 1991; Kesel et al., 1992; Nittrouer et al., 1995; Wright and Nittrouer, 1995; Kuehl et al., 1997; Alexander and Simoneau, 1999; Nittrouer, 1999; Shi et al., 2003; Crockett and Nittrouer, 2004). The highest values are found for rivers that discharge negatively buoyant plumes, due to unusually high sediment concentrations, which rapidly decelerate and deposit sediment near the river mouth (i.e. the Huanghe; Bornhold et al., 1986; Alexander et al., 1991; Wright and Nittrouer, 1995; Shi et al., 2003). The subaqueous deltas of the Amazon River and Ganges-Brahmaputra River, both located in a mid-shelf position (40-70 m of water depth), are the sinks for ~50% and 30% of the annual sediment discharges of each system (Kuehl et al., 1986; Allison et al., 1995; Nittrouer et al., 1995; Wright and Nittrouer, 1995; Kuehl et al., 1997) while it is estimated that 9-15% of the Huanghe sediment leaves the Gulf of Bohai and feeds the subaqueous delta located in the Yellow Sea (Bornhold et al., 1986; Alexander et al., 1991; Wright and Nittrouer, 1995).

Other mid- and outer-shelf (> 70 m water depth) estimates range from 25% to 60% (Wright and Coleman, 1974; Wright and Nittrouer, 1995; Kuehl et al., 1997; Alexander and Simoneau, 1999; Nittrouer, 1999; Sommerfield and Nittrouer, 1999); whereas, the only estimate of sediment sequestered on the upper slope is approximately 20% determined from the Eel River (Alexander and Simoneau, 1999; Nittrouer, 1999). Some rivers at their current highstand positions contribute sediment to submarine troughs and canyons (e.g., Kuehl et al., 1997; Kineke et al., 2000; Mullenbach et al., 2004; Huh et al., 2006). The Okinawa Trough, located off the coast of Taiwan accounts for 5% of the total sediment discharge from Taiwanese Rivers (Huh et al., 2006) while the Eel Canyon collects roughly 12% of the Eel River's annual sediment discharge (Mullenbach et al., 2004).

These budgets do not include sediment transported alongshore via littoral currents or sediment remobilized by waves, currents, or slope-failure mechanisms (Wright and Nittrouer, 1995). The amount of sediment transported alongshore is highly dependent on the energy of the receiving basin manifested through waves and currents as well as the grain size of the suspended load (Wright and Nittrouer, 1995). The Amazon and Eel Rivers discharge to the Atlantic and Pacific oceans, both of which are very energetic regimes. The Amazon River, which carries 90% silt and clay (Wright and Nittrouer, 1995), has been documented to transport at least 17% of its load northward (Kuehl et al., 1986; Allison et al., 1995; Nittrouer et al., 1995); whereas, the Eel River, which carries 75% silt and clay (Sommerfield and Nittrouer, 1999), is thought to loose nearly half of its suspended load farther seaward, landward, or along the shelf (Crockett and Nittrouer, 2004). The remobilization of sediment, on the other hand, can either be significant or negligible. For example the Changjiang deposits an estimated 60% of its load near the river mouth, yet erosion of the sea bed during winter storms removes the majority of the sediment leaving approximately 33% of the original deposit behind as resultant currents transport the eroded sediment to the south where it is redeposited in Hangzhou Bay (McKee et al., 1983; Wright and Nittrouer, 1995). On the opposite end of the spectrum, Mississippi River deposits are believed to remain near the area of initial deposition due to minimal bottom current velocities and low bed stresses (Adams et al., 1987; Wright and Nittrouer, 1995), except in the cases of slope failure and storm events (Adams and Roberts, 1993).

In summary, sediment budgets calculated for river systems are highly dependent on temporal as well as spatial variability (Wright and Nittrouer, 1995) and contain inherent uncertainty due to the upstream locations of gauging stations from the river mouths, which may under- or overestimate the amount of sediment delivered to the coastal ocean (Kuehl et al., 1986). Dispersal systems of ocean basins are affected by sea-level changes, which alters coastal positions and drainage areas of contributing basins, as well as climate changes, which alter storm pathways and intensities. Only taking into account rivers that disperse sediment via positively buoyant plumes, the largest percentage of sediment discharged to the ocean each year is deposited on the continental shelf (Wright and Coleman, 1974; McKee et al., 1983; Bornhold et al., 1986; Kuehl et al., 1986; Alexander et al., 1991; Kesel et al., 1992; Allison et al., 1995; Nittrouer et al., 1995; Wright and Nittrouer, 1995; Kuehl et al., 1997; Alexander and Simoneau, 1999; Nittrouer, 1999; Sommerfield and Nittrouer, 1999; Shi et al., 2003; Crockett and Nittrouer, 2004; Mullenbach et al., 2004; Gebhardt et al., 2005).

These budgets are calculated from accumulation rates and are averaged over decades to years throughout the Holocene and should be used only as a reference. Sommerfield (2006) found that stratigraphic completeness, which is defined as accumulation history including sets of sediment and hiatuses, at the 10^3-year level or resolution varies dramatically depending on the type of system with completeness ranging from 20-39% for deltaic shelves, such as the Amazon shelf, and 51-68% for accretionary shelves, such as the northern California shelf. He also found that: (a) most records are stratigraphically incomplete; (b) completeness increases with water depth; (c) completeness decreases with increases in instantaneous sedimentation rate; and (d) accumulation rates converge at the 10^4-year level of resolution. When comparing the percentage of sediment accumulated within the systems discussed above, the potential for stratigraphic incompleteness must also be considered since records may range from only 20% complete to 68% complete. This stratigraphic completeness could, in part, explain the amount of sediment that remains unaccounted for by the sediment budgets.

Climate and Sea-Level Change

This thesis focuses solely on the last glacial-interglacial cycle. It is common to refer to time periods within the late Pleistocene and Holocene by reference to marine oxygen isotope stages (hereafter MIS; Lambeck et al., 2002), which are based on analysis of the isotopic signatures of foraminifera tests in deep-sea sediment. Variation in the oxygen isotope composition of ocean waters is controlled primarily by the amount of water stored as ice on land (Shackleton and Opdyke, 1973; Lambeck et al., 2002), hence the oxygen isotope record compiled by analyzing foraminifera in deep-sea sediment cores provides an approximation of changes in ice volume and the corresponding glacioeustatic component of sea-level change (Shackleton, 1987).

In terms of MIS, the last glacial cycle is divided into five stages (MIS 1-5). The last interglacial period is included in MIS 5 that lasted from ca. 130,000 to 75,000 years BP (Fig. 6). As summarized by Lambeck and Chappell (2001), on a global scale, MIS 5 was a time of low ice volumes, warm temperatures, and sea level at a highstand position. Increased ice volume, lower temperatures, and lower sea level characterize the conditions during MIS 4 (Guiot et al., 1989; Lambeck et al., 2002), which lasted from ca. 75,000 to 60,000 years BP (Lambeck et al., 2002; Siddall et al., 2003). A decrease in ice volume accompanied by a rise in global sea level occurred during MIS 3, which lasted from ca. 60,000 to 30,000 years BP (Shackleton, 2000; Lambeck et al., 2002; Waelbroeck et al., 2002), while temperatures continued to cool with varying magnitude across the globe (Guiot et al., 1989; Stute et al., 1992; Grimm et al., 1993). The time period from ca. 30,000 to 19,000 years BP, MIS 2, is also known as the last glacial maximum (LGM) because it contains the lowest sea levels (-120 m) at any time during the last glacial cycle due to the large amounts of ice sequestered on land (Lambeck et al., 2002;

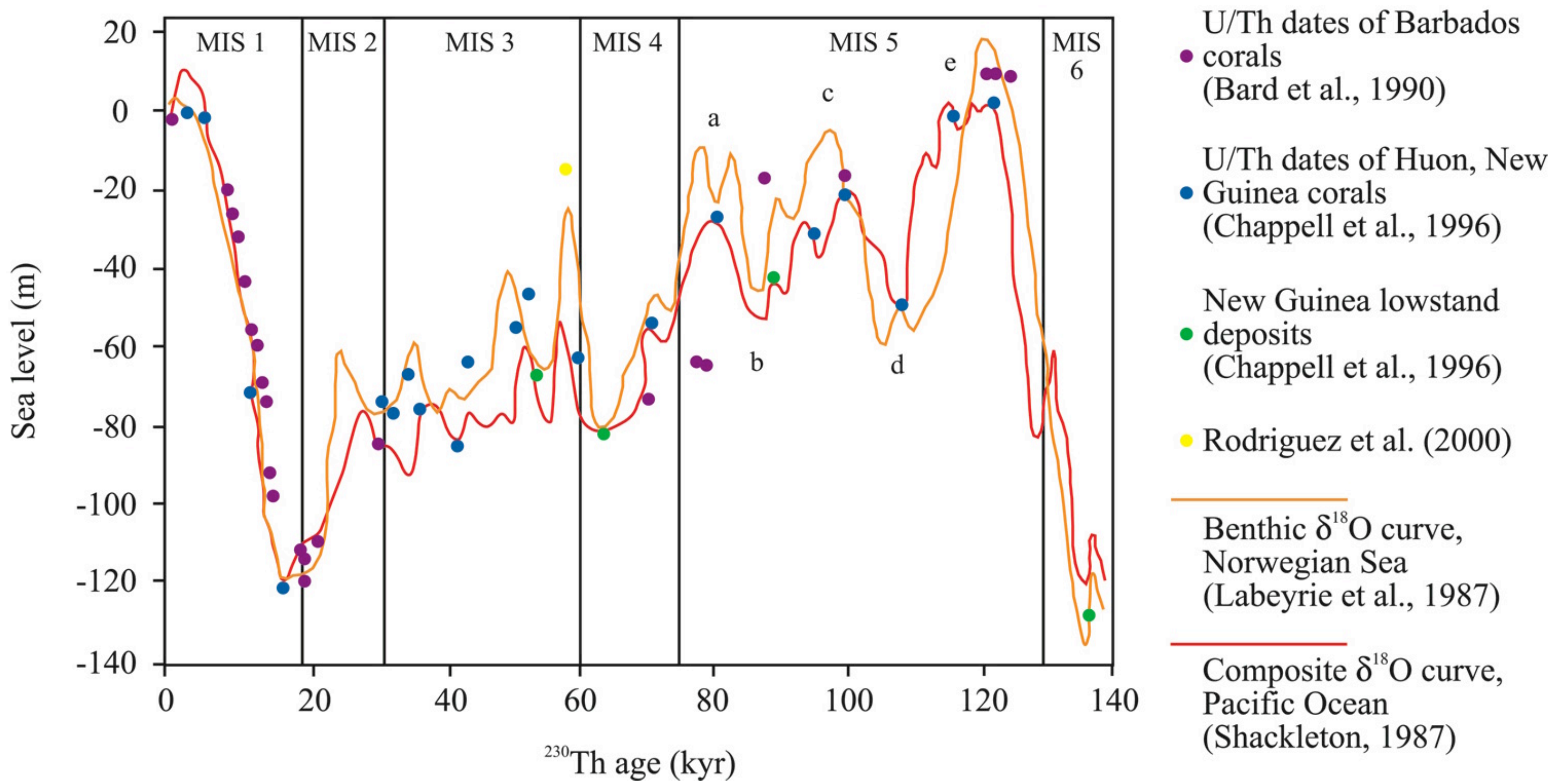


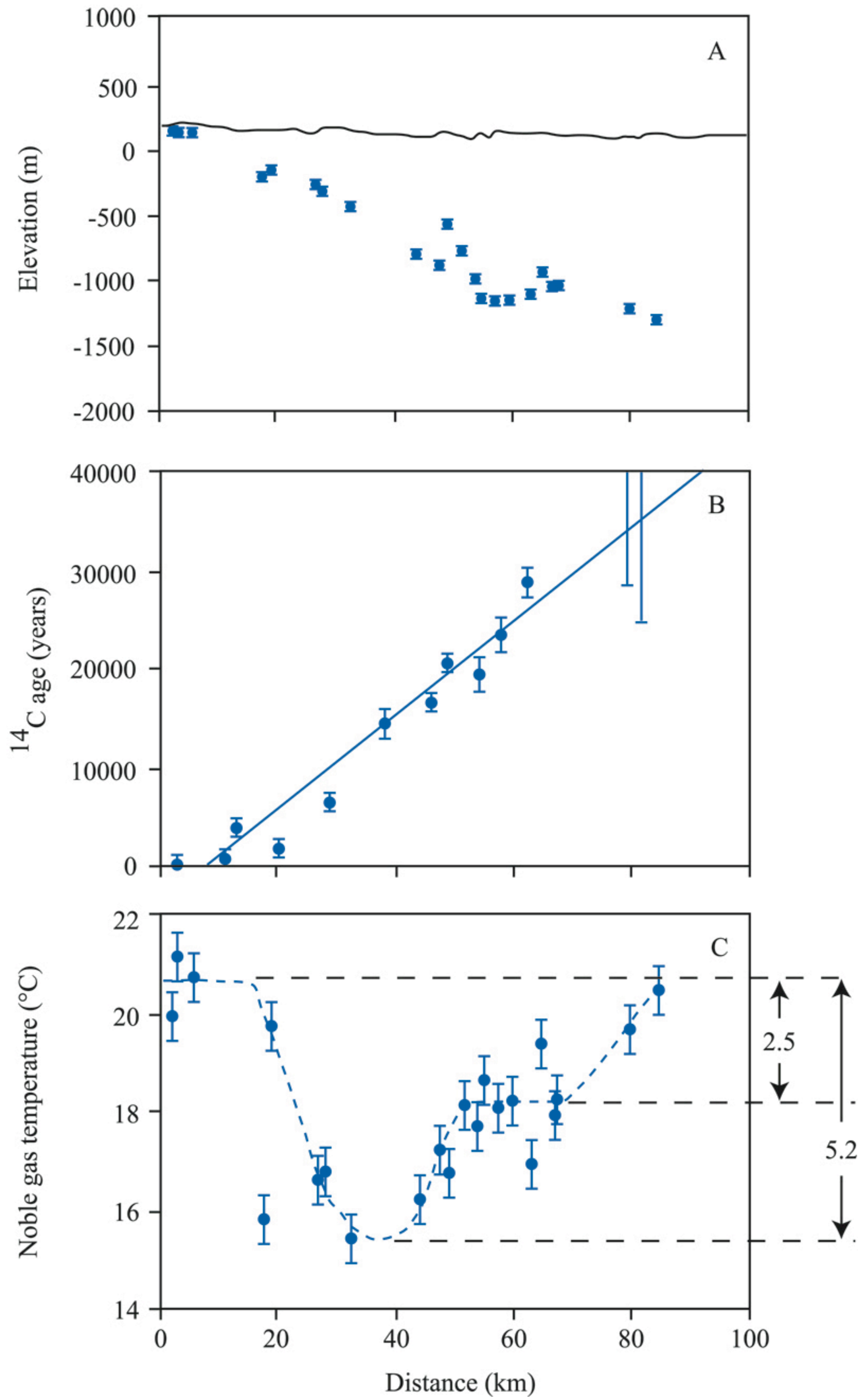
Figure 6. Composite sea-level curve. Developed from composite benthic oxygen isotope curve from the Norwegian Sea (Labeyrie et al., 1987), composite benthic/planktonic oxygen isotope curve from the Pacific Ocean (Shackleton, 1987) adjusted using U/Th dates from Barbados corals (Bard et al., 1990) and Huon, New Guinea corals (Chappell et al., 1996) and converted to sea level elevations. Also included are the sea level datums of New Guinea lowstand deposits (Chappell et al., 1996) and the location of the Texas MIS 3 paleoshoreline position (Rodriguez et al., 2000). Modified from Anderson et al. (2004).

Waelbroeck et al., 2002; Peltier and Fairbanks, 2006). The present interglacial is represented by MIS 1 and has occurred from ca. 19,000 to 0 years BP (Lambeck et al., 2002; Waelbroeck et al., 2002). On a global scale this is a time of decreasing ice volumes (Shackleton, 2000; Lambeck et al., 2002; Waelbroeck et al., 2002) and warming temperatures (COHMAP, 1988; Toomey et al., 1993).

For the northern GOM region, previous researchers have used lake levels, marine microfossils, fossil vertebrate fauna, pollen, plant macrofossil, and speleothem data to investigate the empirical record of paleoclimates (Grimm and Jacobson, 1992; Grimm et al., 1993; Toomey et al., 1993; Stute et al., 1993; Musgrove et al., 2001). Few records exist for MIS 4 and 3, but temperatures were likely cooler than present throughout the region: the only empirical data are provided by Stute et al. (1992) for southwest Texas, near the boundary between the drainage basins of the Colorado River and Rio Grande, who indicate temperature depressions of ~2.5°C during MIS 3 (Fig. 7). For the LGM, ca. 20 kyrs BP, the influence of the Laurentide ice sheet dominated climate of North America, and temperature regimes, precipitation regimes, and the abundance and distribution of plant and animal communities were fundamentally different than today. For the southcentral United States, COHMAP models are in good agreement with empirical estimates for depression of mean annual temperatures by 5-6°C relative to today (Toomey et al., 1993; Stute et al., 1993; Webb et al., 1993; Grimm et al., 1993; Musgrove et al., 2001). Stute et al. (1993) note good agreement with temperature depressions found elsewhere. Similar data are not available for the southeastern United States.

For the southcentral United States a number of different data sources converge to indicate more effective moisture, which is precipitation minus evaporation, several times during MIS 3, and during the LGM (e.g. Toomey et al., 1993; Musgrove et al., 2001). However, the extent to

Figure 7. Paleotemperatures in the southwestern United States derived from dissolved noble gases in the ground water of the Carrizo aquifer (Texas). (A) Vertical section along the Carrizo sandstone recharge flow path with surface topography and well screen locations within the aquifer. (B) ^{14}C ages increase linearly as a function of distance from the recharge area. (C) Changes in noble gas concentrations converted to noble gas temperature as a function of distance from the recharge area, which serves as a measure of age based on the flow velocity determined from (B). This resulted in an estimate of a MIS 3 temperature depression of 2.5°C and a LGM temperature depression of 5.2°C. Modified from Stute et al. (1992).


A
B
C
Elevation (m)
14C age (years)
Noble gas temperature (°C)
Distance (km)
2.5
5.2

which this reflects increased precipitation or decreased evaporation rates due to depressed temperatures is not known. Model estimates of annual precipitation range from close to 40% lower than modern (Webb et al., 1993) to values close to or greater than modern-day precipitation for the southcentral United States (COHMAP, 1988; Toomey et al., 1993; Hall and Valastro, 1995; Musgrove et al., 2001), and as much as 20% lower than present in the southeastern United States (COHMAP, 1988). Studies suggest the southcentral United States would not have experienced convectional storms common to the region today because of lower summer temperatures, and tropical storms were unlikely because of decreased sea surface temperatures (e.g., COHMAP, 1988; Toomey et al., 1993).

Sea Level

Proxy sea-level curves (Fig. 6) have also been established through U-Th dating of *in situ* corals in uplifted reefs of Barbados (Bard et al., 1990) and New Guinea (Chappell et al., 1996). By assuming that global sea level was approximately 6 m above present during the stage 5 interglacial, a mean uplift rate can be calculated, and applied to each dated coral terrace to derive a sea-level curve (Bard et al., 1990; Chappell et al., 1996). Globally, MIS 5 is broken down into five substages with sea levels ranging from ~6-8 m above modern-day sea level (Bard et al., 1990; Chappell et al., 1996) during MIS 5e to as much as 60 m below modern (Lambeck et al., 2002) during MIS 5b. Sea level during MIS 4 continued to decrease with values ranging from 40 m to almost 75 m below present (Chappell et al., 1996; Lambeck et al., 2002; Siddall et al., 2003). A decrease in ice volume caused a temporary rise in sea level during MIS 3 to roughly 40 m below modern (Bard et al., 1990; Chappell et al., 1996; Shackleton, 2000; Lambeck et al., 2002; Waelbroeck et al., 2002). The LGM witnessed a global average sea-level fall to an elevation estimated to be about 120 m below present (Bard et al., 1990; Chappell et al., 1996;

Shackleton, 2000; Waelbroeck et al., 2002; Peltier and Fairbanks, 2006). On the other hand, MIS 1 is characterized by a rapid sea level transgression until around 7,000 years BP (Chappell et al., 1996; Lambeck et al., 2002) followed by a decrease in the rate of sea level rise to reach the present sea level elevation. Figure 6 depicts the proxy sea-level curve for the past 400,000 years with detailed curves for the past 120,000 years.

Within the GOM estimations of sea-level elevations have been derived using radiocarbon dates from shallow-water fauna (e.g. Curray, 1960; Frazier, 1974) and high-resolution seismic stratigraphy (e.g. Abdulah, 1995; Banfield, 1998; Snow, 1998; Anderson, 2001). Records of sea level for the GOM are not as precise as records from elsewhere, but there is no reason to believe there should be significant departures from global eustatic curves, at least at a scale relevant to this thesis. Periods of MIS 5 sea-level fluctuations from +5 m to -40 m were identified in the GOM, though not in all fluvial deposits (Abdulah, 1995; Banfield, 1998; Snow, 1998; Abdulah et al., 2004; Banfield and Anderson, 2004). During MIS 4 it is estimated that sea level reached 80 m below present (Banfield and Anderson, 2004), though MIS 4 deposits are not well preserved within the GOM due to erosion that occurred at the MIS 4 to 3 transition (Anderson et al., 1996; Abdulah et al., 2004; Banfield and Anderson, 2004; McKeown et al., 2004; Roberts et al., 2004). Sea-level elevations during MIS 3, however, range anywhere from present sea level (Frazier, 1974) to -40 m (Abdulah, 1995; Banfield, 1998; Snow, 1998). A recent study by Rodriguez et al. (2000) revised previous MIS 3 estimates from the Texas shelf (Abdulah, 1995; Banfield, 1998; Snow, 1998) to an elevation of -15 ± 2 m based on cores across an escarpment located on the inner shelf. The maximum lowstand elevation ranges from -85 m (Frazier, 1974; Anderson, 2001; McKeown et al., 2004) to the accepted global value of -120 m (Curray, 1960; Abdulah, 1995; Banfield, 1998; Abdulah et al., 2004; Banfield and Anderson, 2004; Roberts et al., 2004). There is complete agreement between the GOM estimates and global estimates for the rate of sea-level rise during MIS 1 to present elevations.

Previous Work on Shelf and Shelf-Margin Deltas

As noted in the introduction, the starting point for this thesis consists of sediment volumes calculated by previous workers. The discussion below provides a brief overview of data used to generate estimated sediment volumes.

Database

Data published in the Anderson and Fillon (2004) volume include over 14,000 km of high-resolution seismic data, which covers most of the northern GOM outer-continental shelf and upper slope, and were collected at various times from the mid 1970s to the late 1990s. Seismic-stratigraphic analysis of each survey followed the methodology established by Mitchum et al. (1977), which uses the configurations and terminations of seismic reflections to identify and correlate depositional sequences, infer depositional environment, and assess lithofacies. Principles of sequence stratigraphy, established by Vail et al. (1977a; 1977b), Posamentier et al. (1988), and Posamentier and Vail (1988) were used to describe and interpret stratal geometries in terms of flooding surfaces, erosional surfaces, and surfaces that separate offlapping and onlapping units.

The flooding surfaces identified by previous researchers are called maximum flooding surfaces, which represent deposition during times when the shoreline is at a maximum landward position (Posamentier and Allen, 1999). There are two types of erosional surfaces that aid in the identification of depositional sequences. The first is called a sequence boundary, which is represented by an erosional unconformity and their correlative disconformities (Vail et al., 1977). A sequence boundary is formed as sea level falls, subaerially exposing sediment from the underlying sequence, which is then eroded and transported basinward (Posamentier and Vail, 1988). Significant erosion occurs as a sequence boundary is formed; whereas, varying magnitudes of erosion and/or nondeposition can accompany the second type of erosional surface called a transgressive surface (Posamentier and Allen, 1999). A rapid rise in sea level creates accommodation at a rate such that sediment supply cannot keep pace thus creating a transgressive surface (Posamentier and Allen, 1999). These surfaces aid in the separation of sediment into depositional sequences, which are then correlated to lithologic descriptions from cores and platform borehole cuttings, when available, to better constrain seismic facies interpretations, as well as estimate the average seismic velocity. Paleobathymetric and paleoenvironmental interpretations are based on the comparison of planktonic and benthic foraminifera from cores to modern living distributions in the GOM.

A combination of oxygen isotope date, planktonic foraminiferal zonations, and ^{14}C ages was used to develop chronological models for the shelf and shelf-margin deltas of offshore Texas, and for the Lagniappe delta of offshore Louisiana and Mississippi (Abdulah et al., 2004; Banfield and Anderson, 2004; Fillon et al., 2004; Fraticelli, 2004). Chronologic constraints for the Alabama and west Florida shelf are less well developed due to the lack of cores or platform borings. McKeown et al. (2004) related the depth of the updip pinch-out and the offlap break to the sea-level proxy curve in order to estimate the timing of accumulation and hence the age of the various units deposited by the Apalachicola system during the last sea-level cycle. Geochronological interpretations made by Correa-Lafayette (2001) are based on correlations between seismic reflectors and a core collected from the Main Pass lease area, within the Mississippi delta region, 70 km to the west of the study area itself. Regardless of chronological limitations, these data are suitable for calculations of minimum values for sediment sequestered in shelf and shelf-margin deltas during the last glacial period.

Previous Calculations of Sediment Volumes in Shelf and Shelf-Margin Deltas

The Apalachicola delta

Anderson (2001) and McKeown et al. (2004) have examined the Apalachicola delta (Fig. 1). Multiple units were identified on the basis of key seismic reflectors and clinoform geometries. Only one unit, Unit 2 of McKeown et al. (2004), is of interest here. It is bounded below by the MIS 5 maximum flooding surface and above by the MIS 2 sequence boundary. Chronological control is based on the relation between the updip pinchout (-60 m) and the offlap break (-90 m) of the unit to the sea-level proxy curve. The unit was interpreted to represent 45 km^3 of sediment by Anderson (2001) and 10.7 km^3 of sediment by McKeown et al. (2004), both of which state the time frame of deposition from 75,000 years BP to 25,000 years BP.

The eastern and western deltas

Sager et al. (1999), Correa-Lafayette (2001), Bart and Anderson (2004), and Bartek et al. (2004) have studied the eastern and western deltas (Fig. 1). Multiple units were identified on the basis of key seismic reflectors and clinoform geometries. Only one unit, Unit 2 of Correa-Lafayette (2001), is of interest here. It is bounded below by the MIS 2 sequence boundary and above by the MIS 1 transgressive surface. Chronological control is based on a seismic correlation established with the core MP303 within the Main Pass lease block, west of the depocenter. The unit was interpreted to represent a combined area of 90 km^3 of sediment, deposited from 75,000 years BP to 13,000 years BP (Correa-Lafayette, 2001) or 18,000 years BP to 16,000 years BP (Bartek et al., 2004).

The Lagniappe delta

The Lagniappe delta (Fig. 1) has been extensively studied. Studies by Kindinger (1988; 1989), Kindinger et al. (1989), Sydow (1992), Sydow et al. (1992), Sydow and Roberts (1994), Roberts et al. (2000), Bartek et al. (2004), and Roberts et al. (2004) were utilized for this thesis. Multiple units were identified on the basis of key seismic reflectors and clinoform geometries. Only one unit, Unit Pro_{10} of Sydow and Roberts (1994), is of interest here. It is bounded below by the MIS 5 maximum flooding surface and above by the MIS 2 sequence boundary. Chronological control is based on AMS ^{14}C ages from cores (Roberts et al., 2004). The unit was interpreted to represent 180 km^3 of sediment (Sydow, 1992; Roberts et al., 2000), deposited from 23,000 years BP to 19,000 years BP (Roberts et al., 2004).

The Brazos delta

Abdulah (1995), Fraticelli and Anderson (2003), Fraticelli (2004), and Abdulah et al. (2004) have examined the Brazos delta (Fig. 1). Multiple units were identified on the basis of key seismic reflectors and clinoform geometries. Only one unit, Unit 3 of Fraticelli (2004), is of interest here. It is bounded below by the MIS 5 maximum flooding surface and above by the MIS 2 sequence boundary. Chronological control is based on ^{14}C ages from cores and platform borings. The unit was interpreted to represent 86 km^3 of sediment, deposited from 44,400 years BP to 33,720 years BP (Fraticelli and Anderson, 2003; Fraticelli, 2004).

The Brazos-Trinity slope system

Badalini et al. (2000), Beaubouef and Friedmann (2000), Beaubouef et al. (2003), and Mallarino et al. (2006) have examined the Brazos-Trinity slope system (Fig. 1). This series of minibasins and their fill were identified on the basis of seismic reflectors. Chronological control is based on 15 giant piston cores taken within minibasin 4 (Mallarino et al., 2006). The volume of these minibasins has not been estimated, but they are thought to represent deposition from 115,000 years BP to 15,000 years BP (Badalini et al., 2000; Mallarino et al., 2006).

The Colorado delta

Abdulah (1995), Snow (1998), van Heijst et al. (2001), and Abdulah et al. (2004) have examined the Colorado delta (Fig. 1). Multiple units were identified on the basis of key seismic reflectors, clinoform geometries (Abdulah, 1995; Snow, 1998; Abdulah et al., 2004) and compiled isopach maps (van Heijst et al., 2001). Only two units, the Stage 3 and Stage 2 deltas of van Heijst et al. (2001), are of interest here. The MIS 3 delta is bounded below by the MIS 5 maximum flooding surface and above by the MIS 2 sequence boundary, whereas the MIS 2 delta is bounded below by the MIS 2 sequence boundary and above by the MIS 1 maximum flooding surface. Chronological control is based on ^{14}C ages from cores and platform borings. The first unit was interpreted to represent 21 km^3 of sediment, deposited from 40,000 years BP to 23,000 years BP by van Heijst et al. (2001) and 10 km^3 of sediment, deposited from 60,000 years BP to 25,000 years BP by Snow (1998). The second unit was interpreted to represent 77.5 km^3 of sediment, deposited from 23,000 years BP to 11,500 years BP by van Heist et al. (2001).

The Rio Grande delta

Banfield (1998) and Banfield and Anderson (2004) have examined the Rio Grande delta (Fig. 1). Multiple units were identified on the basis of key seismic reflectors and clinoform geometries. Only one unit, Lst 2 of Banfield (1998), is of interest here. It is bounded below by the MIS 2 sequence boundary and above by the MIS 1 maximum flooding surface. Chronological control is based on ^{14}C ages from cores and platform borings. The unit was interpreted to represent 150 km^3 of sediment, deposited from 19,000 years BP to 11,000 years BP (Banfield, 1998; Banfield and Anderson, 2004).

METHODS

Calculation of Sediment Supply Rates

Volumes of sediment sequestered in each shelf and shelf-margin delta, as calculated by Sydow (1992), Banfield (1998), Snow (1998), Correa-Lafayette (2001), van Heijst et al. (2001), Fraticelli (2004), and McKeown et al. (2004), provide the framework upon which this thesis builds. As described above, these authors used calculated volumes and the approximate time frame for deposition to estimate sedimentation rates during the last sea-level fall.

Unfortunately, these calculations do not account for porosity, and therefore cannot be directly compared with measurements or model estimates of sediment loads on modern streams, which are in units of mass per unit time. Accordingly, volumes in shelf and shelf-margin delta packages examined in this study are first reduced by 40%, a minimum value, to account for pore waters then converted to units of mass of sediment through dividing by a density of 2700 kg/m^3, a value that represents a mixture of sediment (Meckel et al., 2006) typical of deltaic and shallow marine facies. Supply rates are then calculated using available chronological controls.

Model Estimations of Pre-dam Modern Sediment Supply Rate

In previous studies, comparisons between sedimentation rates from shelf and shelf-margin deltas were compared with sediment supply measured from modern streams (e.g. Banfield, 1998; Snow, 1998; Correa-Lafayette, 2001; Abdulah et al., 2004; Banfield and Anderson, 2004; McKeown et al., 2004). This approach produces a potential source of error, since the sediment discharge values used were strongly biased towards the period after dams were constructed on many streams, and therefore significantly underestimates natural pre-dam sediment yield. In the absence of reliable pre-dam data, an alternative approach would be to estimate modern pre-dam sediment supply using empirical models.

This thesis applies the BQART model to estimate sediment discharges for modern rivers under pre-dam conditions. Application of BQART requires values for drainage area, maximum relief, lithology, and basin-average temperature, with discharge calculated from the model itself. Drainage areas and relief values can be obtained from published sources. The drainage area for each river system was also calculated using RiverTools, a topographic and river network analysis program. RiverTools was used to analyze GTOPO30 digital elevation models (DEM), which cover the Earth at a fixed-angle pixel size of 30 arcseconds. The calculated value of catchment area was then compared with published studies and maps for comparison. When RiverTools returned unrealistic values for the catchment calculated from erroneous river networks due to insufficient resolution of the drainage networks, areas reported by the USGS (2007), Hovius (1998), or Syvitski and Milliman (2007) were used. The maximum relief within each drainage basin, from sea level to the highest point, was determined through the same means.

The addition of a lithology factor to BQART is an improvement in estimating modern sediment supply. Syvitski and Milliman (2007) used, and provide, compiled maps of global geology to assign dominant basin lithologies. The lithology factor ranges from 0.5 to 3, producing a possible 6-fold variation in sediment discharge due to this single factor. From the Syvitski and Milliman (2007) database, the Apalachicola and Brazos Rivers are considered as basins dominated by soft lithologies with areas of harder rocks and thus given a lithology factor of 1.5. The Mobile and Rio Grande drainages have a mixture of hard and soft lithologies, whereas, the Colorado River flows through large areas of carbonate rocks: each of these drainage basins is assigned a lithology value of one.

Both Hovius (1998) and Syvitski and Milliman (2007) compile basin-averaged temperatures for numerous river systems that include those studied in this thesis. In each case, basin-averaged temperatures fall well within the range of modern mean annual temperatures discussed above. Values of modern mean annual temperature are as follows: 17°C for the Mobile basin, 18°C for the Apalachicola and Colorado drainage basins, 19°C for the Brazos drainage basin, and 15°C for the Rio Grande drainage basin.

Model Estimations of Post-dam Modern Sediment Supply Rate

This thesis also tests the applicability of the BQART model by calculating sediment discharge including human impacts and comparing these numbers with modern observed values. The anthropogenic factor takes into account human disturbance of the landscape and varies form 0.3 to 2 (Syvitski and Milliman, 2007). Values compiled by Syvitski and Milliman (2007) give the Apalachicola and Mobile Rivers an anthropogenic factor of 0.3 while the three rivers in Texas are assigned a factor of 1.

The trapping efficiency from Syvitski and Milliman (2007), however, was not used due to insufficient explanation of calculation methods. Instead the trapping efficiency for each basin was calculated using data from dams located in the farthest downstream location and equations taken from Syvitski et al. (2003). All reservoirs on the rivers considered in this thesis, except one, are considered large reservoirs (>0.5 km^3 storage capacity) that do not transfer sediment through, but rather trap all incoming sediment. Claiborne Lake, located on the Mobile River, has a storage capacity of only 0.12 km^3 thus placing it within the small reservoir category (United States Army Corp of Engineers, 1995). For large reservoirs the trapping efficiency can be calculated using:

$$T_E = 1 - 0.05\left(\frac{Q}{v}\right)^{\frac{1}{2}} \tag{7}$$

where: $Q =$ discharge at the head of the reservoir (km^3yr^{-1})

$$v = \text{volume of the reservoir (km}^3)$$

while the trapping efficiency of small reservoirs is calculated using:

$$T_E = 1 - \frac{1}{1 + 0.0021D\dfrac{C}{W}} \quad (8)$$

where: $C =$ reservoir storage capacity (m^3)

$W =$ drainage above reservoir (km^3)

$D = 0.1$

Water discharge at the head of each reservoir is not available so Equation 5 was used to calculate the discharge to that location using only the contributing area to that dam as input. Values for total drainage area, lithology, basin-averaged temperature, and relief are the same as when pre-dam sediment supply was calculated (see previous section).

Model Estimations of Sediment Supply during the Glacial Period

Sediment yields and discharges are estimated for MIS 3 and MIS 2 using the BQART model for each of the extrabasinal fluvial systems described above. Parameters used to estimate pre-dam sediment discharge were modified to account for different glacial period conditions. The following assumptions and glacial-period boundary conditions are used.

1. Hinterland drainage networks, relief, and lithologies remain unchanged during the relatively short time period of 60,000 years from the beginning of MIS 3 to present.
2. Values for drainage area change in accordance with the likely merging of drainages on the shelf: glacial period drainage areas are therefore calculated to include basin-fringe and intrabasinal systems that likely merged with the larger extrabasinal rivers, even though they discharge separately to the coastal oceans today.
3. Total drainage basin relief does not change due to merging of drainage basins.

4. Total relief does change due to sea-level lowering. Values of 60 m are added to MIS 3 relief for sediment supply estimates, and 120 m to relief for sediment supply estimates for MIS 2.
5. A mean annual temperature change of -2.5°C with respect to modern temperature was used for the MIS 3 calculations of the Brazos and Colorado systems. For MIS 2, an average annual temperature change of -3°C with respect to present temperature was used for the Apalachicola and Mobile systems; whereas, a temperature change of -5°C relative to modern was used for the Brazos, Colorado, and Rio Grande systems based on available paleoclimate data.

As noted above, estimates of past precipitation and discharge values are highly uncertain. Hence, several scenarios were used to test the affect of differing hydrological conditions on sediment yield and sediment discharge during the glacial period. Estimated differences in sediment yields are used to illustrate the sensitivity of yield to climate change alone, whereas estimated differences in sediment discharges illustrate the sensitivity to climate and sea-level fluctuations, with corresponding changes in drainage areas and relief. Calculated values of integrated discharge are then compared to estimated masses of sediment sequestered in shelf and shelf-margin deltas.

RESULTS

Present Conditions – Observed vs. Pre-dam Modeled

For the Apalachicola River system, model estimates of pre-dam modern sediment discharge are 10.6 Mt/yr (Table 1), compared to the value of 1.5 Mt/yr cited by Isphording (1986), a number 7 times the modern.

For the Mobile River system, model estimates of pre-dam modern sediment discharge are 11.8 Mt/yr (Table 1), compared to the value of 2.3 Mt/yr cited by Syvitski and Milliman (2007), a number 5.1 times the modern.

For the Brazos River system, model estimates of pre-dam modern sediment discharge are 24.6 Mt/yr (Table 1), compared to the value of 30 Mt/yr cited by Kanes (1970) for a pre-dammed Brazos, a number 0.82 times the modern. The model estimates of pre-dam modern sediment discharge are 2.7 times the 9.1 MT/yr cited by Syvitski and Milliman (2007) for the post-dam Brazos.

For the Colorado River system, model estimates of pre-dam modern sediment discharge are 13.7 Mt/yr (Table 1), compared to the value of 13 Mt/yr cited by Hovius (1998) for a pre-dammed Colorado, a number 1.05 times the modern. The model estimates of pre-dam modern sediment discharge are 9.8 times the 1.4 MT/yr cited by Syvitski and Milliman (2007) for the post-dam Colorado.

For the Rio Grande system, model estimates of pre-dam modern sediment discharge are 106 Mt/yr (Table 1), compared to the value of 30 Mt/yr cited by Hovius (1998) for a pre-dammed Rio Grande, a number 3.5 times the modern. The model estimates of pre-dam modern sediment discharge are 5.3 times the 20 MT/yr cited by Syvitski and Milliman (2007) for the post-dam Rio Grande.

Table 1. Model inputs and results using modern conditions.

River	Area (km^2)	Relief (km)	Temp (°C)	Glacial (%)	Lithology	1 - T_E	Erosion	Q_s (MT/yr)	Yield ($T/km^2/yr$)
Apalachicola	50,000	1.245	18	0	1.5	1.00	1.00	10.6	212
Mobile	113,500	1.188	17	0	1.0	1.00	1.00	11.8	104
Sabine Neches	51,550	0.145	19	0	1.5	1.00	1.00	1.34	26
Trinity San Jacinto	56,800	0.140	19	0	1.5	1.00	1.00	1.39	24.5
Brazos	117,425	1.441	19	0	1.5	1.00	1.00	24.6	210
Colorado	109,400	1.344	18	0	1.0	1.00	1.00	13.8	126
Rio Grande	471,900	4.198	15	0	1.0	1.00	1.00	107	227

Present Conditions – Observed vs. Post-dam Modeled

For the Apalachicola River system, roughly 90% of the drainage area lies behind the farthest downstream dam. Model estimates of post-dam modern sediment discharge, with a trapping efficiency of 70%, are 0.8 MT/yr. Compared to the value of 1.5 MT/yr calculated by Isphording (1986), the model estimates are 0.5 times the modern (Table 2).

For the Mobile River system, roughly 50% of the drainage area lies behind the farthest downstream dam. Model estimates of post-dam modern sediment discharge, with a trapping efficiency of 31%, are 2.4 MT/yr. Compared to the value of 2.1 MT/yr cited by Judson and Ritter (1964), the model estimates are 1.1 times the modern (Table 2).

For the Brazos River system, almost 40% of the drainage area lies behind the farthest downstream dam. Model estimates of post-dam modern sediment discharge, with a trapping efficiency of 80%, are 5.6 MT/yr. Compared to the value of 9.1 MT/yr cited by Syvitski and Milliman (2007), the model estimates are 0.6 times the modern (Table 2).

For the Colorado River system, approximately 90% of the drainage area lies behind the farthest downstream dam. Model estimates of post-dam modern sediment discharge, with a trapping efficiency of 82%, are 2.5 MT/yr. Compared to the value of 1.4 MT/yr cited by Syvitski and Milliman (2007), the model estimates are 1.8 times the modern (Table 2).

For the Rio Grande system, roughly 80% of the drainage area lies behind the farthest downstream dam. Model estimates of post-dam modern sediment discharge, with a trapping efficiency of 80%, are 21.4 MT/yr. Compared to the value of 20 MT/yr cited by Syvitski and Milliman (2007), the model estimates are 1.07 times the modern (Table 2).

Table 2. Model inputs and results using modern conditions including human impacts compared to modern observed values taken from Judson and Ritter (1964), Isphording (1986), and Syvitski and Milliman (2007).

River	Area (km^2)	Relief (km)	Temp (°C)	Glacial (%)	Lithology	$1 - T_E$	Erosion	Q_s (MT/yr) Modeled	Q_s (MT/yr) Observed
Apalachicola	50,000	1.245	18	0	1.5	0.25	0.30	0.80	1.5
Mobile	113,500	1.188	17	0	1.0	0.68	0.30	2.41	2.3
Brazos	117,425	1.63	19	0	1.5	0.20	1.00	5.57	9.1
Colorado	109,400	1.36	18	0	1.0	0.18	1.00	2.51	1.4
Rio Grande	471,900	4.198	15	0	1.0	0.20	1.00	21.4	20

Test Cases – Modeled Changes of Sediment Supply

The sediment discharge in kilograms per second was calculated for each system first using modern values (Table 1) then using four different scenarios to illustrate sensitivity of sediment supply to different controls. The first scenario holds drainage area constant and varies the basin-average temperature above and below present values to simulate sediment yields due to fluctuations in temperature (Fig. 8). The second scenario holds basin-average temperature constant, but increases drainage basin area to full glacial values (Fig. 9), which illustrates how sediment supply changes in response to merging and/or lengthening of drainage basins during sea-level fall and lowstand. For the third scenario, both drainage area and the basin-average temperature were varied in the same manner as above, which illustrates the estimated combined effect of changes in temperature, and merging of drainage basins during the glacial period (Table 3). The fourth scenario has two parts: (a) drainage area and basin-averaged temperature were varied in the same way as scenario 3 while the discharge was increased by 25% and decreased by 25% to simulate wet and dry glacial conditions (Fig. 10) and (b) drainage area and basin-averaged temperature mimic modern conditions while a range of discharges are used to simulate wet and dry modern conditions (Fig. 11). Percent changes outlined below represent the differences between the cases mentioned above and a modeled control value (Table 1), which reflects modern area and temperature with no human influence.

The first scenario varies average annual temperature over a range of temperatures from 30°C to 5°C (Fig. 8) and because drainage area is held constant, this scenario simulates changes in sediment yields due to only temperature change. This range encompasses full glacial temperatures, which are based on temperature depressions of 3° C for the Apalachicola and Mobile River systems, and 5°C for the Brazos, Colorado, and Rio Grande systems. Model

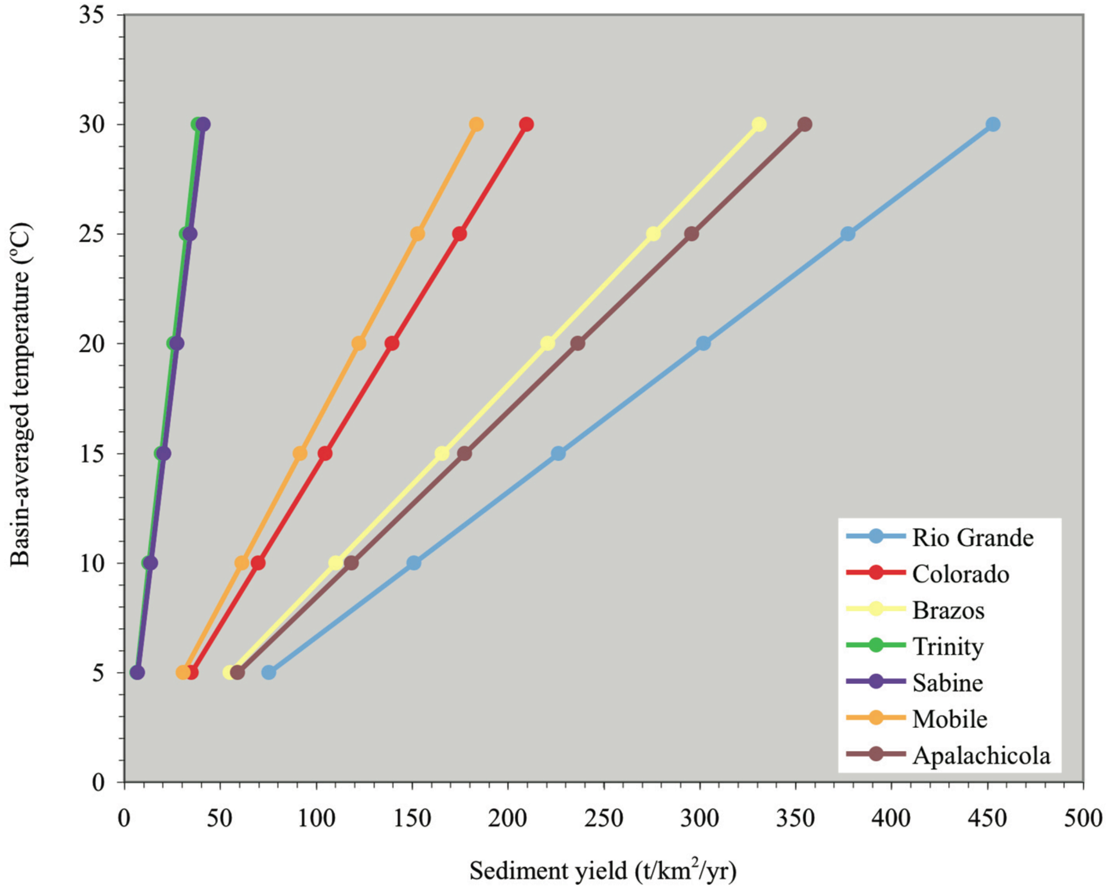


Figure 8. Graph of sediment yield (mass per unit area per unit time) versus basin-averaged temperature for all river systems illustrating changes in sediment yields due to changes in temperature.

estimates for the LGM temperature depressions range from a 17% decrease in yields in the case of the Apalachicola River, to 33% for Rio Grande. The paragraphs below summarize changes due to the other scenarios mentioned above, with all values reported as a percent change from model predictions for the modern interglacial highstand system.

For the Apalachicola system, a 1% increase in drainage area is assumed, based on the extension across the continental shelf and addition of exposed shelf area. Holding temperature constant, a 1% increase in drainage area produces an estimated increase in sediment discharge of 1%. Combining this increase in drainage area with a temperature depression of 3°C results in a decrease in sediment discharge of less than 8% relative to predicted modern values (Table 3).

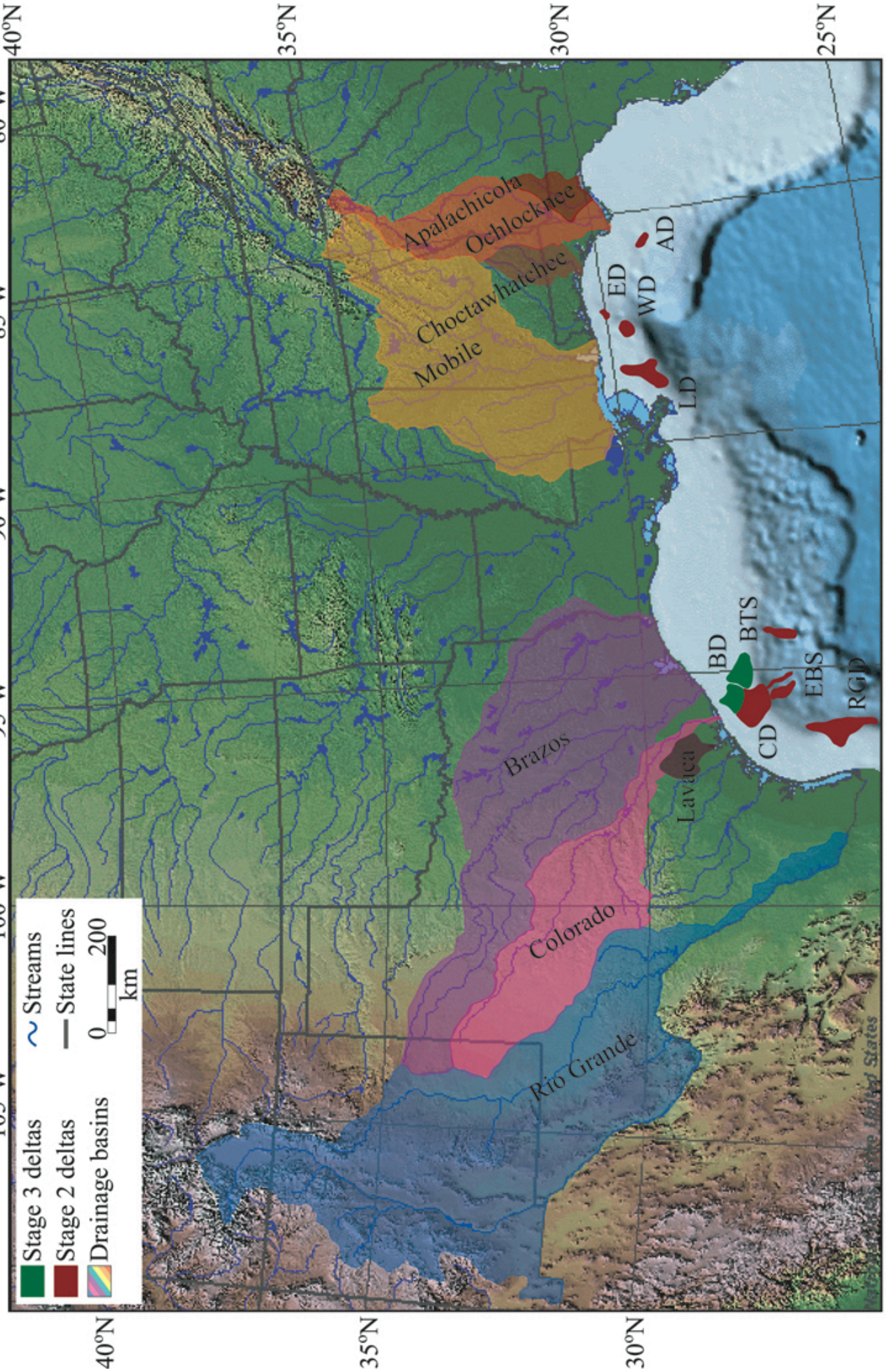
For the Mobile River system, a 46% increase in drainage area (Fig. 9) is assumed, based on the possibility of it merging with the Pascagoula, Pearl, Tangipahoa, Tickfaw, and Amite Rivers as well as extension across the continental shelf and addition of exposed shelf area. Holding temperature constant, a 46% increase in drainage area produces an estimated increase in sediment discharge of 33%. Combining this increase in drainage area with a temperature depression of 3°C results in an increase in sediment discharge of 20% relative to modern values (Table 3).

For the Brazos River system, a 110% increase in drainage area (Fig. 9) is assumed, based on merging with the Trinity and Sabine Rivers (Abdulah, 1995; Beaubouef and Friedmann, 2000; Fraticelli, 2004) as well as the extension across the shelf and addition of exposed shelf area. Holding temperature constant, a 110% increase in drainage area produces an estimated increase in sediment discharge of 65%. Combining this increase in drainage area with a temperature depression of 5°C results in an increase in sediment discharge of 30% relative to predicted modern values (Table 3).

Table 3. Model inputs and results using glacial conditions.

River	Area (km^2)	Relief (km)	Temp (°C)	Glacial (%)	Lithology	$1 - T_E$	Erosion	Q_s (MT/yr)	Yield ($T/km^2/yr$)
Apalachicola	50,650	1.365	15	0	1.5	1.00	1.00	9.82	194
Mobile	166,160	1.308	14	0	1.0	1.00	1.00	14.2	85.5
Brazos Trinity Sabine	246,375	1.561	14	0	1.5	1.00	1.00	34.2	139
Colorado	116,500	1.46	13	0	1.0	1.00	1.00	11.3	96.9
Rio Grande	472,795	4.318	10	0	1.0	1.00	1.00	73.4	155

Figure 9. Location map showing shaded relief, drainage basins thought to be present during the last glacial maximum, deltas associated with each, and general bathymetry. The different deltas are labeled as follows: AD – Apalachicola delta (McKeown et al., 2004), ED – eastern delta (Bart and Anderson, 2004), WD – western delta (Bart and Anderson, 2004), LD – Lagniappe delta (Roberts et al., 2000), BD – Brazos delta (Fraticelli, 2004), BTS – Brazos-Trinity slope system (Beaubouef et al., 2003), CD – Colorado delta (van Heijst et al., 2001), EBS – east breaks slide (van Heijst et al., 2001), RGD – Rio Grande delta (Banfield and Anderson, 2004). Base map modified from National Atlas of the United States (2007).


Stage 3 deltas
Stage 2 deltas
Drainage basins
Streams
State lines
0 200
km
Apalachicola
Ochlocknee
Choctawhatchee
Mobile
AD
ED
WD
LD
BD
BTS
EBS
RGD
CD
Brazos
Lavaca
Colorado
Rio Grande
40°N
35°N
30°N
25°N
80°W
85°W
90°W
95°W
100°W
105°W

For the Colorado River system, a 6% increase in drainage area is assumed, based on the likely occurrence of it merging with the Lavaca River as well as extension across the shelf and addition of exposed shelf area. Holding temperature constant, a 6% increase in drainage area produces an estimated increase in sediment discharge of 4%. Combining this increase in drainage area with a temperature depression of 5°C results in a decrease in sediment discharge of 18% relative to predicted modern values (Table 3).

For the Rio Grande system, a 0.2% increase in drainage area is assumed, based on the extension across the shelf and addition of exposed shelf area. Holding temperature constant, a 0.2% increase in drainage area produces an estimated increase in sediment discharge of less than 1%. Combining this increase in drainage area with a temperature depression of 5°C results in a decrease in sediment discharge of 33% relative to predicted modern values (Table 3).

The last scenario simulates the effects of both increases and decreases in precipitation for each river system. Because of the exponent in BQART, increasing or decreasing discharge always produces a non-linear change (Fig. 11) in predicted sediment load. However, for all river systems, increasing water discharge by 25% under glacial-boundary conditions as well as modern produced an increase in sediment discharge of 6-8% (Figs. 10 and 11). When water discharge was decreased by 25% for modeled glacial conditions (Fig. 10) as well as modeled modern conditions (Fig. 11), a decrease in sediment discharge of 8-9% was produced for all river systems.

Sediment Volumes – Model and Observed Estimates

The shelf-margin delta interpreted to represent the Apalachicola River (Fig. 1, 9) during the last glacial period contains an estimated volume of 10.7 km^3: correcting for porosity and converting to units of mass yields a value of 17,300 Mt. Previous workers estimate this delta

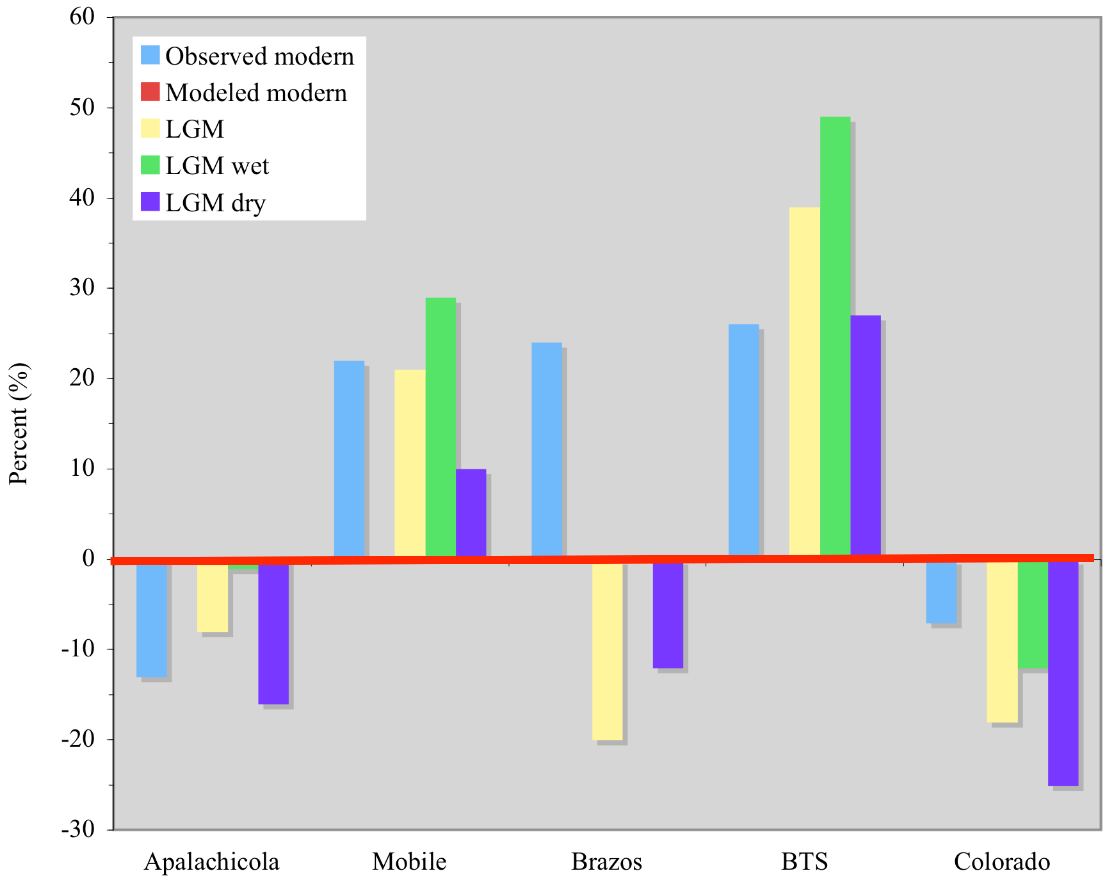


Figure 10. Bar graph illustrating the percent differences in sediment discharge with respect to the modeled-modern sediment discharge.

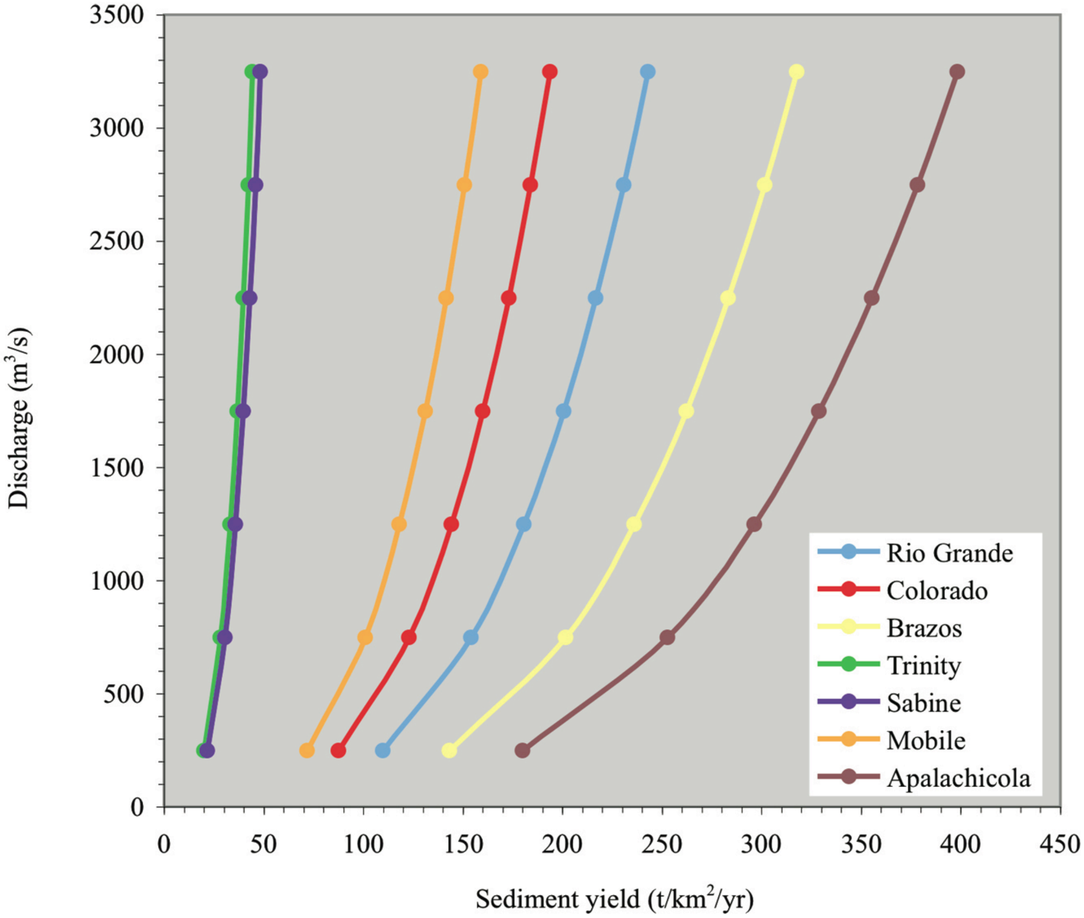


Figure 11. Graph of sediment yield (mass per unit area per unit time) versus water discharge illustrating changes in sediment yield due to changes in precipitation.

was deposited during a period of 50 kyrs, which yields an average rate of deposition of 346 Mt/kyr. Model estimates of sediment discharge for the Apalachicola River, with drainage area increased by 650 km^2 and a temperature depression of 3°C, are 9.81 Mt/yr or 9810 Mt/kyr. With estimated rates of this magnitude, sediment mass observed within the shelf-margin delta would have been produced within 1.76 kyrs or over a longer time frame if sediment dispersal dynamics were such that less of the sediment budget was used to build the delta. Alternatively, integrated over the inferred time period of deposition, model estimates would result in 490,500 Mt of predicted sediment discharge unaccounted for.

The shelf-margin deltas interpreted to represent the Mobile River (Fig. 1, 9) during the last glacial period contain estimated sediment volumes of 90 km^3 and 180 km^3: correcting for porosity and converting to units of mass yields values of 145,800 Mt and 291,600 Mt. Previous workers estimate these deltas were deposited during periods of 55 kyrs and 4 kyrs, which yields average rates of deposition of 2650 Mt/kyr and 72,900 Mt/kyr. Model estimates of sediment discharge for the Mobile River, with drainage area increased by 51,600 km^2 (Fig. 9) and a temperature depression of 3°C, are 11.8 Mt/yr or 11,800 Mt/kyr. With estimated rates of this magnitude, sediment mass observed within the shelf-margin deltas could have been produced within 12.3 kyrs and 24.7 kyrs or over longer time frames if sediment dispersal dynamics were such that less of the sediment budget was used to build the delta. Alternatively integrated over the inferred time periods of deposition, model estimates would result in 503,200 Mt of predicted sediment discharge unaccounted for and a deficit of sediment produced of 244,400 Mt.

The shelf-margin delta interpreted to represent the Brazos River (Fig. 1, 9) during the last glacial period contains an estimated sediment volume of 86 km^3: correcting for porosity and converting to units of mass yields a value of 139,320 Mt. Previous workers estimate this delta was deposited during a period of 11.4 kyrs, which yields an average rate of deposition of 12,220 Mt/kyr. Model estimates of sediment discharge for the Brazos River, with drainage area increased by 1,000 km^2 and temperature depression of 2.5°C during MIS 3, are 22.4 Mt/yr or 22,400 Mt/kyr. With estimated rates of these magnitudes, sediment mass observed within the shelf-margin delta could have been produced within 6.22 kyrs or over a longer time frame if sediment dispersal dynamics were such that less of the sediment budget was used to build the delta. Alternatively, integrated over the inferred time period of deposition, model estimates would result in 116,040 Mt of predicted sediment discharges unaccounted for.

The slope minibasins interpreted to represent the Brazos-Trinity system (Fig. 1, 9) during the last glacial period have not been measured volumetrically. Previous workers estimate these minibasins were filled during a period of 100 kyrs. Model estimates of sediment discharge for the Brazos-Trinity system, with drainage area increased by 128,950 km^2 (Fig. 9) and temperature depression of 5°C, are 32 Mt/yr or 32,000 Mt/kyr. Integrated over the inferred time period of deposition, model estimates would result in 3,200,000 Mt of predicted sediment discharge.

The shelf and shelf-margin deltas interpreted to represent the Colorado River (Fig. 1, 9) during the last glacial period contain estimated sediment volumes of 21 km^3 and 77.5 km^3: correcting for porosity and converting to units of mass yields values of 34,020 Mt and 125,550 Mt. Previous workers estimate these deltas were deposited during periods of 17 kyrs and 11.5 kyrs, which yields average rates of deposition of 1,118 Mt/kyr and 6,000 Mt/kyr. Model estimates of sediment discharge for the Colorado River, with drainage area increased by 2,000 km^2 and 10,920 km^2 (Fig. 9) and temperature depressions of 2.5°C and 5°C, are 12.5 Mt/yr or 12,500 Mt/kyr and 11.3 Mt/yr or 11,300 Mt/kyr. With estimated rates of these magnitudes, sediment mass observed within the shelf and shelf-margin deltas could have been produced within 2.72 kyrs and 11.1 kyrs or over longer time frames if sediment dispersal dynamics were such that less of the sediment budget was used to build the delta. Alternatively, integrated over the inferred time periods of deposition, model estimates would result in 178,480 Mt and 4,400 Mt of predicted sediment discharge unaccounted for.

The shelf-margin delta interpreted to represent the Rio Grande (Fig. 1, 9) during the last glacial period contains an estimated volume of 150 km^3: correcting for porosity and converting to units of mass yields a value of 243,000 Mt. Previous workers estimate this delta was deposited during a period of 8 kyrs, which yields an average rate of deposition of 30,375 Mt/kyr. Model estimates of sediment discharge for the Rio Grande, with drainage are increased by 900 km^2 and temperature depression of 5°C, are 73.4 Mt/yr or 73,400 Mt/kyr. With estimated rates of this magnitude, sediment mass observed within the shelf-margin delta could have been produced within 3.31 kyrs or over a longer time frame if sediment dispersal dynamics were such that less of the sediment budget was used to build the delta. Alternatively, integrated over the inferred time period of deposition, model estimates would result in 344,200 Mt of predicted sediment discharge unaccounted for.

DISCUSSION

Key results with first-order interpretations can be summarized as follows:

- Model estimates of sediment yield for the last glacial period are 9 to 38% lower than modeled present values. The controlling factor appears to be temperature. Model predictions of decreases in sediment yield contrast previous interpretations of greatly increased sediment supply during the glacial period (see Abdulah et al., 2004; Anderson, 2005; Banfield and Anderson, 2004; Banfield, 1998; Correa-Lafayette, 2001; Sydow and Roberts, 1994)).
- In most cases, model estimates for sediment discharge during the last glacial period are also lower than modeled present values, except for the Mobile and Brazos Rivers. The range of predicted decreases in sediment discharge for other extrabasinal streams is 8% for the Apalachicola system to 33% for the Rio Grande system. For the Mobile River, the possible merging of the Pascagoula, Pearl, Tickfaw, Amite, and Comite systems would increase drainage area by 46%: model predictions for sediment discharge to the river mouth are 21% greater than the sediment discharge for the present-day highstand Mobile. For the Brazos River, due to merging of the Trinity and Sabine systems (Abdulah, 1995; Beaubouef and Friedmann, 2000; Fraticelli, 2004), drainage area increased by a 110%: model predictions for sediment discharge to the river mouth are 30% greater than the sediment discharge for the present-day highstand Brazos.
- It is likely that all the extrabasinal systems discussed in this thesis were joined by tributary streams during the last glacial period, as they crossed the shelf. In some cases, increases in drainage area were relatively minor, but in the case of Brazos River, total drainage area during lowstand was likely to be approximately twice that of the present

day. Increases in drainage area result in model predictions of increases in sediment discharge that are proportional to the area increase, as a power-law relation (Eq. 2, 3). If sediment discharge to a river mouth was greater during the last glacial period than it is today, the primary reason would have been merging of drainage areas, rather than increases in sediment yield.

- In all but one case, observed sediment volumes, when corrected for porosity and converted to units of mass, are significantly less than model estimates of sediment discharge with glacial period drainage areas and climate boundary conditions over the inferred time periods of deposition. Put another way, sediment mass observed in all but one of the shelf-margin deltas examined in this study could be produced in significantly less time than the inferred time period of deposition, and a large portion of predicted sediment discharge remains unaccounted for by existing observations. The amount that remains unaccounted for could be attributable to errors of some kind (see below) or the mass of sediment that might be present in slope and basin floor systems farther downdip.
- The lone anomaly is the Lagniappe delta, which contains an observed sediment volume that requires a 30% increase in sediment discharge relative to the modeled glacial discharge for the Mobile without taking into account the sediment beneath the St. Bernard delta or that lost to deep water (Sydow and Roberts, 1992). This increase implies that either: (a) volume estimates are overestimated, (b) other extrabasinal systems contributed, or (c) discharge increased despite the temperature depression.

Sources of Uncertainty

As mentioned before, sea-level records for the GOM are not as robust as records from elsewhere. However, there is no reason to believe there were significant departures from the global eustatic curves, and the only manner by which sea level impacts the BQART calculations is in terms of changes in relief, where the magnitude of uncertainty is small relative to total relief. Climate estimations over the last interglacial to glacial period are sparse for the United States, but exact values of paleoclimate and modern climate are not required due to the fact that the model utilizes basin-averaged temperatures. Drainage area and relief are dependent on the organization of the landscape through tectonic activity and can be considered constant over the last glacial cycle. Therefore sea level, climate, drainage area, and relief are fairly well constrained for the study area over the relatively short time period and do not contribute significant uncertainty.

BQART Model

One warning frequently expressed when using empirical equations of this kind is that the output is only as good as the input. The conundrum, however, lies in the fact that a model's applicability can only be determined by the quality of its results (Olea, 1999). The BQART model utilized in this thesis, when tested on a database of 488 rivers, had a bias of approximately 3% while accounting for 96% of the data variance (Syvitski and Milliman, 2007). This model is a predictive tool and by no means free from error, but more importantly the source of the uncertainty and therefore error can be identified.

Drainage area and relief explain 57% of the between-river variability and do not contribute significant sources of uncertainty over the relatively short period focused on in this thesis (see above). The influx of glaciermelt and snowmelt, on the other hand, increases the variability of sediment yield (Meybeck et al., 2003), but rivers in the study area did not drain glaciated regions during the LGM. Lithology is the other geologic control on sediment production and the factors compiled by Syvitski and Milliman (2007), which reflect the dominant basin lithologies, were used in calculations for this thesis. The exception is the Colorado River that drains large areas of carbonates and is best represented by a lithology factor of 1 instead of the 1.5 reported by Syvitski and Milliman (2007). For larger basins, there is more variability and uncertainty that can be included within the lithology factor because large basins contain a range of lithologies that are not always accurately embodied by one basin-averaged factor. This uncertainty is inherent and can only be reduced by applying factors that represent the average lithology of the basin, which have been consistent over the time period of interest. Hence, any uncertainties associated with the use of specific values for lithology, or more broadly, for all geologic factors, are the same for modern pre-dam, modern post-dam, and LGM estimates.

The BQART model calculates discharge, which explains 3% of the variance, from the basin area (Syvitski and Milliman, 2007). This may seem insignificant, but some of the variance that might otherwise be attributable to discharge is accounted for by the area term because discharge is closely correlated to and predicted by drainage area. Basin-averaged temperature, on the other hand, accounts for 10% of the variance and, like drainage area, is well constrained for the study area.

One exercise of this thesis was to compare observed post-dam values of discharge and sediment load to calculated values, which took into account the human effects on the river system. The goal of this exercise was to test the model's predictive capabilities of modern sediment discharge in order to determine its overall applicability. To do this a trapping efficiency, calculated for the farthest downstream dam, and the anthropogenic factor reported by Syvitski and Milliman (2007) were used. The model estimates ranged form 0.5 to 1.8 times the post-dam sediment discharges. The estimates for all rivers, except one, fall within the cloud of data presented by Syvitski and Milliman (2007). The Colorado River is the only river to fall outside the ± 50% with a model estimate 1.8 times the post-dam observed sediment discharge. The model predicts the maximum amount delivered to the coastal ocean implying that either: (a) observed values are underestimated; (b) that there is a significant amount of sediment retained within the system; or (c) that the model does not accurately estimate sediment discharge. Based on the other rivers within the study area, it is believed that the model does provide estimates of sediment supply comparable to the observed post-dam values and can thus be used to estimate sediment discharge under modern conditions without human influence as well as glacial conditions.

Sources of Error

Sediment Supply Rates

Overall, error associated with the calculation of sediment supply rates can be summarized as follows:

- Estimates of sediment volumes measured within shelf and shelf-margin deltas are underestimations of the total sediment deposited during the last glacial period due to: (a) sediment that bypassed the delta and are deposited on the continental slope, in deep water, or elsewhere in the distributary system cannot be accounted for; (b) depending on the bounding surfaces, sediment eroded during sea-level regression or transgression cannot be accounted for; and (c) sediment sequestered within the floodplain and coastal plain cannot be accounted for.
- Calculations of sedimentation rates during the glacial period are based on the volume of sediment packages, equivalent to filled accommodation. These calculations did not correct for porosity, and therefore cannot be directly compared with measurements or

model estimates of sediment loads on modern streams, which are in units of mass per unit time. This source of error significantly overestimates the mass of sediment delivered, since porosity can be 40-70% in shallow deltaic sediment of the Gulf of Mexico and elsewhere (Morton and Price, 1987; van Heijst et al., 2001; Meckel et al., 2006). The value of 40% chosen for this thesis is a conservative value, so likely overestimates the observed total mass of sediment stored in shelf-margin deltas, but not to the same degree as original estimates that did not account for porosity at all.

- Chronological controls are less than ideal, since much of the record is older than the effective limits of radiocarbon dating (<35,000 yrs), and estimated age ranges may therefore be incorrect. Age control is especially sparse along the eastern portion of the study area. For example, the Apalachicola delta, eastern delta, and western delta do not have cores from which age control could be obtained so depositional periods were estimated from the offlap break position and a distant seismic correlation, respectively.

Another source of error arises when calculated sediment supply rates are compared with sediment supply rates measured from modern streams. The numbers associated with modern streams are strongly biased towards the period after dams were constructed on many streams, and therefore significantly underestimate natural pre-dam sediment yield. Moreover, there is at present no way to ascertain whether values measured for the short period prior to dam construction on some rivers are actually representative, or whether they might represent enhanced periods of soil erosion common to the early half of the 20th Century.

Lagniappe Delta

It can be argued that rivers studied in this thesis fit patterns of sediment supply to and deposition in shelf-margin deltas that make sense within the context of the BQART model approach used here. The lone exception to this general statement would be the Lagniappe delta, which raises a number of interesting questions regarding the lowstand drainage of the area east of Mississippi valley, the largest river of interest here being the Mobile.

The exact location of the deposits left behind by the Mobile River has been debated (Kindinger et al., 1994; Sydow and Roberts, 1994; Sager et al., 1999; Correa-Lafayette, 2001; Bart and Anderson, 2004; Roberts et al., 2004). There is evidence linking the Mobile River, either alone or with a combined drainage including the Pearl and Pascagoula Rivers, to depocenters to the southeast (Sager et al., 1994; Correa-Lafayette, 2001; Bart and Anderson, 2004) and southwest (Kindinger, 1988; Kindinger, 1989; Kindinger et al., 1994; Sydow, 1992; Sydow et al., 1992; Sydow and Roberts, 1994; Roberts et al., 2004) of Mobile Bay.

The southwest depocenter, the Lagniappe delta of Kindinger (1988), has been studied extensively (see Kindinger, 1988; 1989; Sydow, 1992; Sydow et al., 1992; Sydow and Roberts, 1994; Winn et al., 1995; Fillon et al., 2004; Kohl et al., 2004; Roberts et al., 2004). Based on AMS ^{14}C dates obtained from cores and an extensive seismic grid, it was determined that the Lagniappe Delta contains 190,000 Mt of sediment, after accounting for porosity (Sydow, 1992; Roberts et al., 2000), and was deposited between 23,000 yrs BP and 19,000 yrs BP (Roberts et al., 2004).

As previously mentioned it is possible that the Mobile, as it extended across the shelf, merged with the Pascagoula, Pearl, Tangipahoa, Tickfaw, and Amite Rivers to increase its total drainage by 46%, though whatever evidence there may have been for this merging is no longer present. Model estimates of sediment discharge rates, using this increased area and a temperature depression of 3°C, are approximately 11,800 Mt/kyr, which equals only 47,200 Mt of sediment deposited over the 4,000-year period the Lagniappe is interpreted to represent, or roughly 142,800 Mt less than what is contained in the Lagniappe. This estimate does not include sediment lost to the deep basin, eroded during sea-level regression, or hidden beneath the acoustic wipeout region caused by the Holocene St. Bernard delta of the Mississippi River (Sydow et al., 1992). This leads to a series of possible questions: (1) Are sediment volumes overestimated? (2) Is the timing of deposition sufficiently constrained? (3) Is it possible that sediment yields increased despite the 3°C temperature depression? (4) If a larger drainage basin is necessary to deliver the sediment sequestered in the Lagniappe delta, what additional drainage area might that be?

Future Studies

Lagniappe Delta

A number of workers consider the source of the Lagniappe delta to be the Mobile River, perhaps combined with the Pascagoula and Pearl Rivers (Kindinger, 1988; Kindinger, 1989; Kindinger et al., 1994; Sydow et al., 1992; Sydow and Roberts, 1994; Roberts et al., 2004), but without a link to the Mississippi system. Even with a 46% increase in area, model estimates of sediment discharge are 142,800 Mt less than required to form the Lagniappe delta, without accounting for sediment deposited off the shelf edge, eroded during sea-level regression, and hidden by the St. Bernard delta. It is possible that other drainages also contributed sediment to the Mobile during the last glacial period. A first step towards resolving issues of contributing drainage area would be a high-resolution provenance study: such a study would test the inferred Mobile-Pearl-Pascagoula source, and possibly identify other candidate drainage areas.

Brazos-Trinity Slope System

After the Brazos joined the Trinity-Sabine system (Fig. 9), modeled sediment fluxes increased relative to the calculated modern value of the Brazos River alone. A comparison to the fill within the minibasins cannot be made until a calculation of sediment volumes is done. It is believed that deposition within basin 4, which is the seaward extent of the system, began around 115,000 years BP (Mallarino et al., 2006), but at that time it is unlikely that the Brazos was a contributor of sediment. The four minibasins are linked and have accumulated sediment fill-and-spill style since the last interglacial (Badalini et al., 2000). This would indicate that gravity flows generated by sediment delivered to the shelf by the combined Trinity and Sabine, a drainage area as large as the modern Colorado or Brazos, were the first to arrive in the basins. Better age constraints would be helpful in determining the exact timing of deposition of units within the basins. This would allow model inputs to be calibrated to estimate how much sediment could have been deposited. These results could then be used to better understand sediment delivery and transport dynamics within the system.

Sediment Dispersal Systems

Recent studies (e.g., Wright and Coleman, 1974; McKee et al., 1983; Bornhold et al., 1986; Kuehl et al., 1986; Alexander et al., 1991; Kesel et al., 1992; Nittrouer et al., 1995; Wright and Nittrouer, 1995; Kuehl et al., 1997; Kineke et al., 2000; Shi et al., 2003; Crockett and Nittrouer, 2004; Mullenbach et al., 2004; Gerbhardt et al., 2005; Huh et al., 2006) provide estimates of the amount of sediment retained in various parts of a fluvial system. Most studies (e.g., McKee et al., 1983; Kuehl et al., 1986; Alexander et al., 1991; Nittrouer et al., 1995; Kuehl et al., 1997; Kineke et al., 2000; Crockett and Nittrouer, 2004; Mullenbach et al., 2004; Gerbhardt et al., 2005; Huh et al., 2006) have focused on sediment dispersal to offshore or alongshore localities using values of sediment discharge from gauging stations located large distances inland from the river mouth. The amount of sediment stored within the delta plain remains unknown or is estimated to be the sediment supply unaccounted for.

In the case of ancient examples, the amount of sediment sequestered within a delta is known while the amount of sediment lost to other parts of the system remain unknown with no way to quantify the amount lost. It is difficult to relate modern sediment budgets to ancient systems due to conditions that were fundamentally different compared to the present, for example the modern global transgression and highstand, versus the last glacial period falling stage and lowstand. Sea level was 120 m below the modern level and rivers were extended across the shelf to discharge at or near the shelf edge. There have also been studies that suggest rivers at this time were transporting a higher percentage of bedload than at present (e.g., van Heijst et al., 2001; McKeown et al., 2004). This would alter the sediment dispersal pattern significantly by increasing the amount of sediment deposited within a delta.

Future studies of sediment budgets and dispersal systems should include the entire delta plain, seaward of the last gauging station, as well as offshore regions such as the shelf, slope, and deep basin. Deltas discharging to positions at or near the shelf edge could also be used as possible analogs for glacial-period rivers. These sediment budgets could then be used to better understand the dynamics of sediment supply, delivery, and dispersal during glacial periods.

CONCLUSIONS

Workers have suggested that the amount of sediment discharged to the northern GOM was significantly greater during the last glacial period relative to today. This thesis tested the plausibility and necessity of these previous interpretations and attempted to move toward a first-generation sediment budget that contrasts interglacial versus glacial periods for river systems that discharge to the northern GOM. This thesis has not developed predictions that can be tested; rather it distills a couple fundamental relationships, and identifies further questions using recently developed empirical models (Syvitski et al., 2003; Syvitski and Milliman, 2007). Primary conclusions are as follows:

1. Model estimates for sediment yield during the last glacial period are lower than present values due to temperature depressions and do not support previous interpretations that sediment supply greatly increased during the last glacial period. Although this conclusion is derived from an empirical model, it is considered robust, and more broadly applicable than the specific study area here, because of the general positive correlation between sediment supply and temperature.
2. Only two extrabasinal systems discussed in this thesis likely experienced a significant increase in sediment supply during the last glacial period. However, rather than an increase in sediment yield, this increase was due to the merging of drainage basins increasing the area contributing to a single river mouth point source. Increases in drainage area of the Mobile and Brazos Rivers resulted in model predictions of increases in sediment discharge that are proportional to the area increase. Model estimates for sediment discharge of the other major extrabasinal systems during the last glacial are also lower, due to reduced sediment yields, and the smaller increases in drainage area. This

conclusion also seems robust, namely that if specific river systems did experience an increase in sediment discharge, it does not represent an increase in sediment yield, but rather an increase in drainage area due to merging of drainages during lowstand.

3. In all but one case, observed sediment volumes, when corrected for porosity and converted to units of mass, are significantly less than model estimates of sediment discharge with glacial period drainage areas and glacial climate boundary conditions over the inferred time periods of deposition. In all but one case, therefore, a significant percentage of predicted sediment discharge remains unaccounted for, and was presumably available for dispersal elsewhere, perhaps to deepwater.
4. The Lagniappe delta is the only delta that contains more sediment than can be supplied with glacial period drainage areas and climate boundary conditions, and over the inferred time periods of deposition. This problem is exacerbated when considering the volume of sediment that was likely dispersed to deepwater, and not measured within the Lagniappe. The Mobile system during the last glacial period, therefore requires either a significant increase in drainage area or discharge, beyond that attributable to including the Pearl and Pascagoula systems, to produce the amount of sediment present in the Lagniappe delta over a 4,000-year time frame, or the timing of deposition for the Lagniappe system needs to be reevaluated.

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